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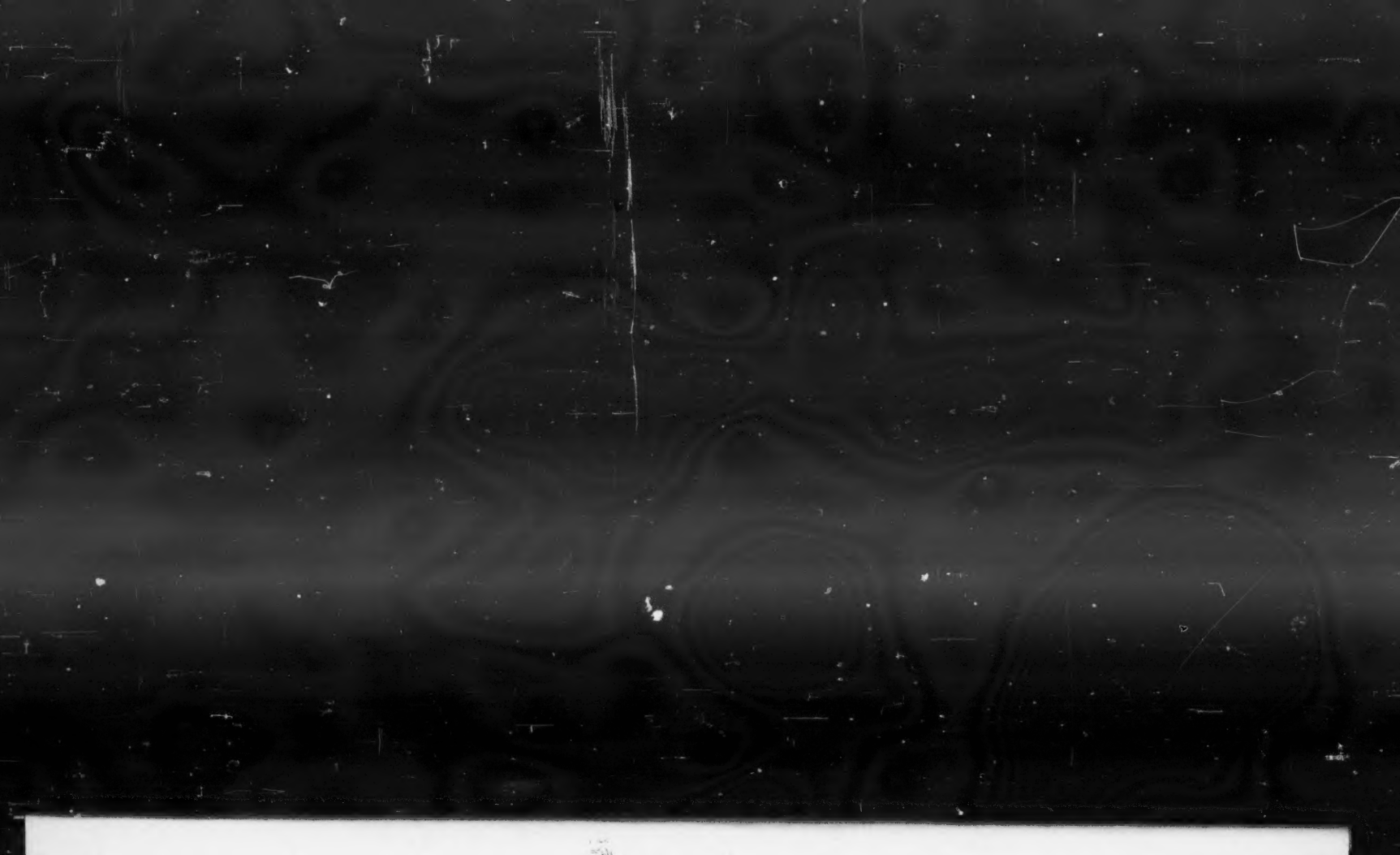

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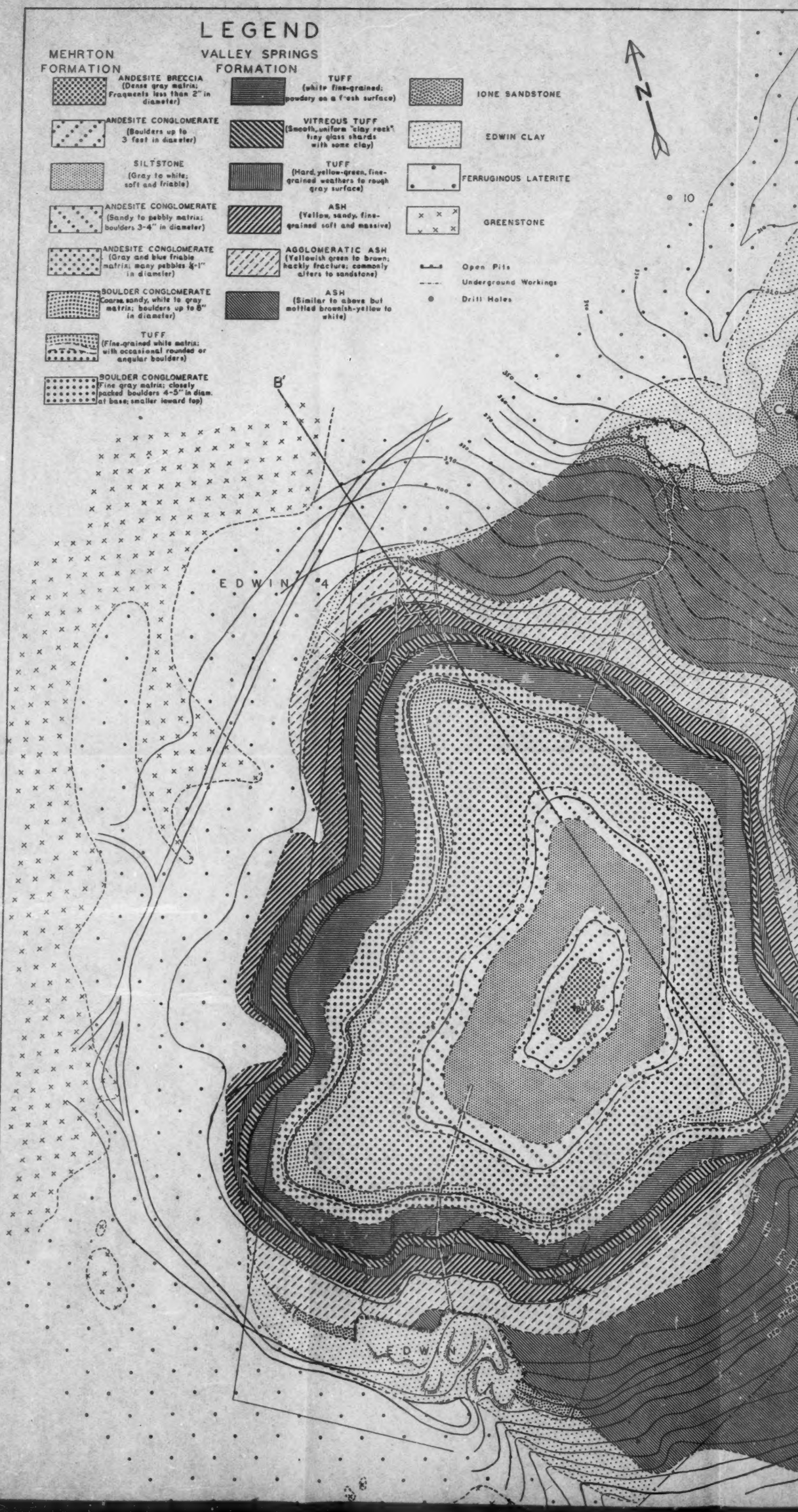
LEGEND
MEHRTON FORMATION
ANDESITE BRECCIA (Dense gray matrix; Fragments less than 2" in diameter)
ANDESITE CONGLOMERATE (Boulders up to 3 feet in diameter)
SILTSTONE (Gray to white; soft and friable)
ANDESITE CONGLOMERATE (Sandy to pebbly matrix; boulders 3-4" in diameter)
ANDESITE CONGLOMERATE (Gray and blue friable matrix; many pebbles ¼-1" in diameter)
BOULDER CONGLOMERATE Coarse, sandy, white to gray matrix; boulders up to 8" in diameter)
TUFF (Fine-grained white matrix; with occasional rounded or angular boulders)
BOULDER CONGLOMERATE Fine gray matrix; closely packed boulders 4-5" in diam. at base; smaller toward top)
VALLEY SPRINGS FORMATION
TUFF (white fine-grained; powdery on a fresh surface)
VITREOUS TUFF (Smooth, uniform "clay rock" tiny glass shards with some clay)
TUFF (Hard, yellow-green, fine-grained weathers to rough gray surface)
ASH (Yellow, sandy, fine-grained soft and massive)
AGGLOMERATIC ASH (Yellowish green to brown; hackly fracture; commonly alters to sandstone)
ASH (Similar to above but mottled brownish-yellow to white)
IONE SANDSTONE
EDWIN CLAY
FERRUGINOUS LATERITE
GREENSTONE
Open Pits
Underground Workings
Drill Holes
N
10
B'
C
EDWIN 4
EDWIN 3

A
A'
C
C'
EDWIN #2
EDWIN #3
B
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Scale
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SILTSTONE
(Gray to white; soft and friable)
ANDESITE CONGLOMERATE
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B'
C
EDWIN 4
USGS
BM
Topography from map by
V. Sullivan and O.M. Tupper

GEOLOGIC MAP OF JONES BUTTE, IONE, CALIFORNIA

LIFORNIA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 56, PP. 1-38, 5 PLS., 5 FIGS. JANUARY 1945

ORIGIN OF THE EDWIN CLAY, IONE, CALIFORNIA

BY THOMAS FULCHER BATES

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ABSTRACT

The Edwin clay forms a commercially important deposit 3 miles west of Ione, California. It is a sedimentary clay of the Ione formation of Eocene age.

In the type locality the Ione sediments lie in a narrow trough between the Sierra Nevada source rocks on the east and an outlying ridge of greenstone on the west. Investigation of the clay and of accompanying quartz grains shows that the Ione clays are separable into two mineralogical groups, the Cheney Hill clays along the western side of the trough, and the Eastern Ione clays in the center and on the eastern side.

Detailed studies of inclusions in the quartz grains place the Edwin clay in the Cheney Hill group, although it contains less quartz than other clays of this type.

Electron micrographs, X-ray-and electron-diffraction patterns, chemical analyses, differential thermal analyses, firing tests, and fluorescent tests with morin dye show that the Edwin clay consists of kaolinite with appreciable amounts of gibbsite, hematite, and authigenic anatase, in addition to quartz grains and a suite of heavy minerals.

The rocks of the Sierra Nevada provided the clay, quartz, and heavy minerals. Ferruginous laterite, residual upon the greenstone, furnished the gibbsite and hematite. Much of the hematite stain in the clay was removed by organic solutions produced under stagnant-water conditions. The presence of CO_2 in the solutions caused much of the iron to be precipitated in the lower part of the clay as siderite.

INTRODUCTION

Ione has long been one of the chief ceramic clay centers of California. Producing pits are numerous, and much exploration has been conducted in search for additional deposits. Clay strata, with beds of sand and lignite, occur in the Ione formation of Eocene age. The clays have been used for pottery and for fire brick. Although they vary considerably in color, plasticity, and other physical properties changing abruptly within short distances, as a whole they exhibit characteristics which group them as a sedimentary unit.

One variant, the Edwin clay, however, shows distinctive features. Although on testing it shows the high shrinkage and fire cracking typical of the Ione clays, it is more refractory and has long been recognized as one of the best California plastic fire clays.

The present study was undertaken to ascertain the geological and mineralogical reasons for the more obvious differences between the Edwin clay and the other clays of the Ione formation and to explain, in more detail than has heretofore been done, the origin of this exceptional variety.

ACKNOWLEDGMENTS

This study has been made possible by the kind co-operation and generosity of the Stockton Fire Brick Company[1] through its president, Mr. John T. Roberts. Mr. Vernon Sullivan, ceramic engineer, Mr. O. M. Tupper, mining engineer, and Mr. Charles Butts, mine superintendent, have been of great assistance. Information on the clays and general geology of the region freely made available by Mr. A. B. Wallen, clay prospector, was of great aid in the study of the many clay and sand pits and of numerous drill holes around Ione. Considerable laboratory assistance has been furnished by the company's research laboratory at Pittsburg, California, and for this help I am sincerely indebted to Mr. G. A. Page and Mr. F. F. Raine.

The thermal analyses were made by the author with the equipment of the Department of Ceramics of the Pennsylvania State College through the kindness of Dr. E. C. Henry.

The electron micrograph and the electron-diffraction pattern of the clay were taken at the Princeton laboratory of the Radio Corporation of America through the courtesy of Dr. E. W. Engstrom and Dr. James Hillier.

Throughout this entire study Dr. Paul F. Kerr of the Department of Geology, Columbia University, has rendered most valuable assistance in both field and laboratory work. It was at his suggestion that this work was undertaken and through his guidance and encouragement that it has been pursued and completed. Mr. Ralph J. Holmes, of the same department, has given unsparingly of his time, particularly in helping with some of the X-ray work.

I am further indebted to Dr. Paul D. Krynine of the Department of Earth Sciences of the Pennsylvania State College for many valuable suggestions especially with respect to the study of the quartz inclusions.

GENERAL CONSIDERATIONS

LOCATION

Ione is in Amador County, California, at the junction of the Sacramento Valley with the foothills of the Sierra Nevada 33 miles southeast of Sacramento and 33 miles northeast of Stockton (Fig. 1).

The environs have long been of geological importance because of the sands and lignite which, together with the clays, constitute the Ione formation. The former were once mined for commercial purposes.

Iron ore has commonly been mentioned (Whitney, 1865; Irelan, 1888; Turner, 1894) in connection with the Ione formation due to an occurrence west of Ione of highly ferruginous laterite stratigraphically beneath the Ione formation. Attempts were once made to mine the material.

[1] The Stockton Fire Brick Co. has recently been taken over by Gladding, McBean and Co., Los Angeles, California.

Ione lies 1½ miles west of the edge of the Sierra Nevada province in the California Trough section. The Trough section in the latitude of Ione has been divided (Piper, Gale, Thomas, and Robinson, 1939, p. 15) into four subsections: the Delta country or Delta tidal plain, the Victor alluvial plain, the river flood plains and

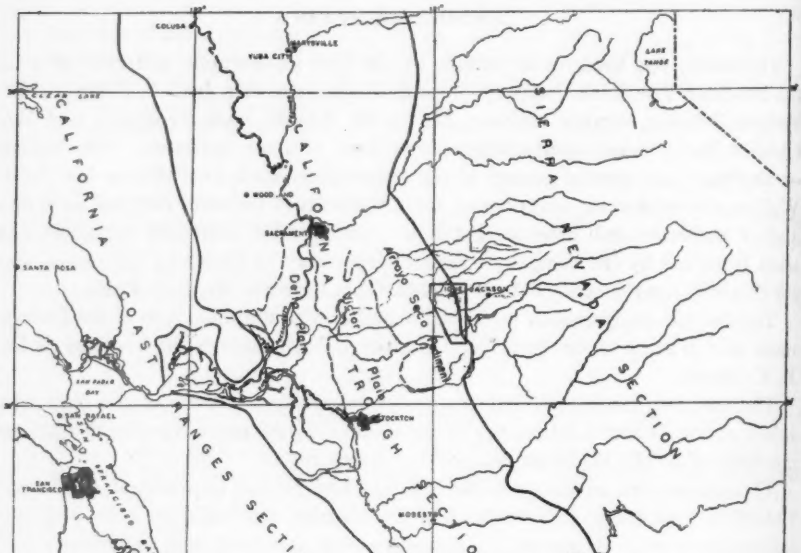


FIGURE 1.—*Index map of central California*
(After Piper, Gale, Thomas, and Robinson, 1939)

channels, and the Arroyo Seco dissected pediment. Ione lies in the dissected pediment in a broad depression trending somewhat west of north and separating the Sierra Nevada foothills on the east from the few remaining pediment remnants on the west. The floor of this depression is irregular due to differential erosion, and consequently the definition of this topographic feature varies locally. It is most pronounced west and southwest of Ione where its western edge is defined by a discontinuous, low, elongate ridge which parallels the edge of the Sierra Nevada foothills but lies 2½ to 3 miles west of them. This ridge is formed by two outliers of the so-called bedrock series of the Sierra Nevada and is of major importance in the problem of the origin of the Edwin clay.

Even in the best-developed part of the depression there are prominent irregularities. Most obvious of these is Jones Butte, 3½ miles west of Ione; this prominence rises sharply over 350 feet above the valley floor on the east, 100 feet above the ridge immediately west of it, and nearly as high above the elevation of the projected pediment.

There are three distinct geologic units in the vicinity of Ione which are of importance in this discussion (Fig. 2):

(1) East of Ione the foothills of the Sierra Nevada are underlain by closely folded greenstones, argillites, and slates of the bedrock series along the western edge of this mountain range. Two outliers of greenstone, already mentioned because of the ridge they form, lie west of Ione and appear to form a link between similar rocks exposed south of the Calaveras River and the greenstone found beyond the Cosumnes River to the north. These rocks are generally considered to be Jurassic and Carboniferous.

(2) Resting unconformably upon this series are the sands and clays of the Ione formation. These are products of long weathering which have been transported from their original position in the Sierra Nevada and deposited in deltas in a shallow marine embayment. This arm of the sea spread farther east in the vicinity of Ione than it did to the north or south due to a trough which isolated the outlying greenstone ridge from the rest of the bedrock series. The sediments in this trough dip moderately westward and are of particular interest because of their high content of the clay mineral, anauxite (Allen, 1928).

(3) Unconformably overlying the Ione formation and also resting locally upon the older rocks is an assemblage of detrital sediments consisting of rhyolitic and andesitic tuff, ash, and agglomerate. Except at Jones Butte where it forms a resistant cap and just north of the butte where it laps around the northern edge of the exposed greenstone, this volcanic detritus has been eroded away between the greenstone outliers and the Sierra Nevada foothills, exposing the underlying Ione formation. Lindgren (1911) proposed that the source of the volcanic materials was 60 miles or more east of the western edge of the foothills. The rocks have been separated by Piper *et al.* (1939) into the Valley Springs and Mehrton formations and are thought to be Miocene and Pliocene.

POSITION OF THE CLAY DEPOSITS

The general area of Ione deposition in the vicinity of Ione is a narrow trough parallel to the Sierra Nevada, which bounds it on the east, and to the low ridge of outlying greenstone which forms the western edge (Fig. 2). The Cheney Hill variety of Ione clay lies along the western side of the area, farthest from the eastern source.

Deposit No. 1 (Fig. 2), the Edwin clay deposit, lies beneath the resistant volcanics which form Jones Butte. The clay has been extensively worked in open pits and in underground operations. These workings, supplemented by test holes, outline an area of about 120 acres.

Deposit No. 2, at Smoky Corners, consists of several acres of clay locally called the Edwin type. The deposit has been prospected by augur borings.

Deposit No. 3, on the side of Cheney Hill, gives this type of clay its name.

Deposit No. 4 is at the Woolford pit, while the clay of deposit No. 5 lies in China Gulch and is known by that name.

Occasional deposits farther east may contain a bed of clay of the Cheney Hill

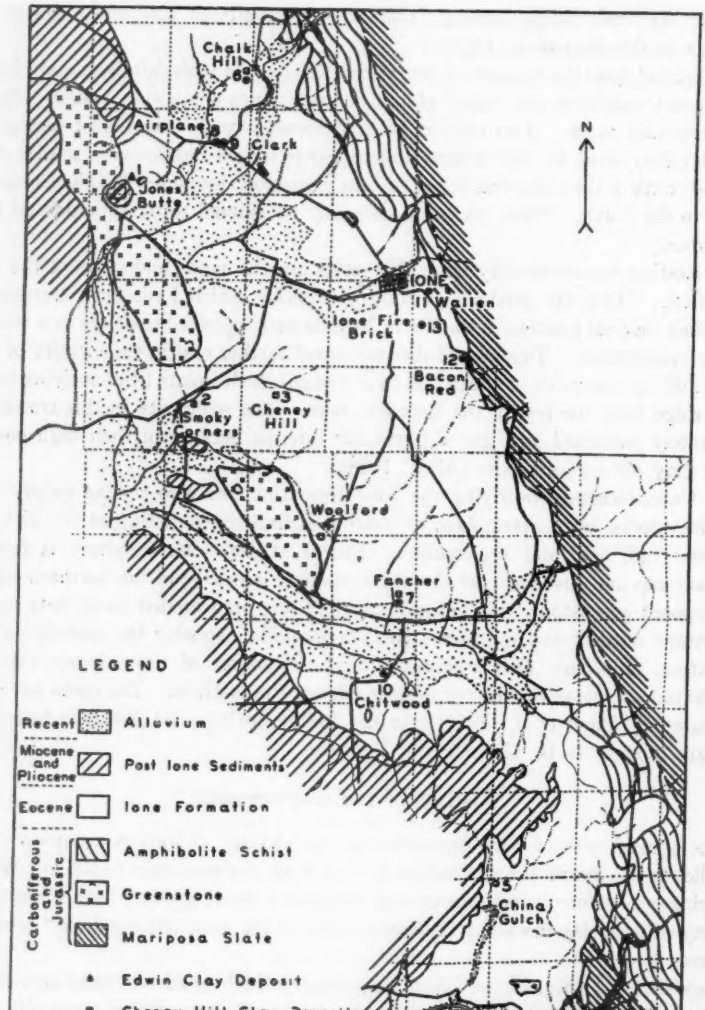


FIGURE 2.—*Geologic map of the Ione area, California*
(Modified from Piper, Gale, Thomas, and Robinson, 1939)

type although most of the material belongs to the Eastern Ione clay type. Such deposits are No. 6 at Chalk Hill and No. 7 at the Fancher pit.

The other clay deposits (Nos. 8–13) consist of the typical Eastern Ione clay. Location No. 13 includes 8 or 10 closely connected pits.

FORMER STUDIES OF THE IONE FORMATION

Two modes of origin have been proposed for the Ione clays. Lindgren (1894, p. 3) mentions both possibilities in his description in the Sacramento folio where the term Ione formation was first used to describe a "sedimentary series consisting of clays and sands, . . . in places there are indications that the clays have been derived from rhyolitic tuffs." Allen (1929) has shown that the latter concept was due to the original inclusion within the Ione formation of part of the overlying series of rhyolite tuffs. These have locally been altered to form the "clay rock" described by Turner (1894) with which the older sedimentary clays of the Ione formation, as redefined by Allen, have been confused.

> "Statements have appeared from time to time that the lower clays of the Ione formation were formed from rhyolitic tuffs. The investigation of the nature of these clays lends no support to this notion, but on the contrary shows that both the clays and the associated sands were derived from a surface that had been subjected to a long period of chemical weathering" (Allen, 1929, p. 354).

This contention is supported by the work in the area of Piper *et al.* (1939), who have placed the "clay rock" in the overlying Valley Springs formation and have divided the Ione formation into three members (p. 80):

> "These are an upper member of massive white sandstone; an intermediate member that includes alternating lentils of white or light colored clay and beds of sand similar to the upper member; and a lower member that consists of gray or bluish shale and clay, lignite or coal, and other carbonaceous beds."

This view suggests that the clays of the Ione formation are sedimentary in origin and that the rhyolitic tuffs are unconformable upon the Ione sands and clays. Detailed studies herein recorded confirm this conclusion.

FORMER STUDIES OF THE EDWIN CLAY

Most of the State reports on clay mining near Ione do not mention the Edwin clay as such nor distinguish between the clay mined at Jones Butte and the other clays of the Ione formation. Thus, Logan (1927, p. 135) speaks of the deposit operated by the Stockton Fire Brick Company and gives the following section from the pit at Jones Butte:

> "3 feet..... Soil and andesite boulders increasing uphill
> 10 feet..... White and iron-stained sand and sandy clay
> 10–14 feet..... White fire clay
>
> Red clay"

No indication is given of the distinguishing characteristics of this clay, an understandable omission because without firing tests the Edwin clay may easily be mistaken for one of the more typical varieties of Ione clay.

Dietrich (1928) records the results of firing tests made on samples from all the operative pits in the State. He stated (p. 272) that the Edwin clay "is one of the best of the Ione fire clays" and further (p. 54) "one of the best plastic fire clays in California."

The geologic relations described in Dietrich's paper were those worked out by C. N. Schuette; his description of the geology of the deposit and his interpretation of its origin were later published (1929). Schuette inferred no connection between the Edwin clay and the Ione clays but stated (p. 196) that "the clay deposit is a limited portion of the clay tuff formation" which lies above it. As to the mode of origin, Schuette continues as follows (p. 197):

> "The clay is an alteration product of the clay tuff and was formed practically in place by kaolinizing action of meteoric waters and humus acids. Brown iron oxide still found in the inner portions of the clay bed was probably carried into the originally white tuff from the underlying laterite by magmatic waters."

Allen (1929, p. 381) makes no reference to this possible mode of origin for the Edwin deposit. Under the subheading "Ione clays" he mentions the Edwin clay as follows:

> "At Jones Butte is another type of clay which might be regarded as a further stage of decomposition. A white brittle clay, locally known as the 'Edwin clay' forms the lowest eight feet of the deposit. Its plasticity is lower than the usual Ione clay and it breaks with a conchoidal or irregular fracture."

In another section (p. 404) he suggests that the clay at Jones Butte came from the upper part of the weathering profile of the bedrock to the east and for that reason "has a lower proportion of fluxes than . . . beds which were derived from a lower less-leached part of the profile." In addition he states:

> "The fine clays were flocculated by their contact with the salt water; and shifting currents at times mixed grains of sand with clay, producing unusual types such as the 'Edwin clay,' and the clay with quartz grains near Carbondale."

Thus, Allen recognized differences which distinguish the Edwin from the typical Ione clay and indicated they may be due to "a further stage of decomposition" and a shifting of currents which mixed sand grains with the clay. He shows that the Edwin clay was derived, like the clays of the Ione formation, by weathering and erosion of the older rocks of the Sierra Nevada, transportation westward, and deposition as delta deposits in a shallow Eocene sea.

Field work during the summer of 1941 led the author to publish an abstract (Bates, 1942) stating that the Edwin is a residual clay formed by alteration of the underlying laterite. Subsequent laboratory studies have proved this incorrect; instead they verify Allen's conclusions as to the sedimentary origin of the material, although the author's observations do not bear out Allen's suggestion that the mixing of sand grains with the much finer-grained Edwin clay is one of the features which makes it unusual. Such a mixture is typical of all the Ione clays of the Cheney Hill type.

The present study is an attempt to add to these general concepts more detail as to the nature and origin of the several types of Ione clay near the town of Ione and of the Edwin clay in particular.

GEOLOGY OF JONES BUTTE

The largest and best-developed deposit of Edwin clay lies on the east slope of the greenstone ridge beneath the younger sediments which form Jones Butte (Pl. 1).

North, west, and south of the center of the Butte the greenstone is exposed. East of the hill drill holes strike this rock at depths of 80 to 100 feet, indicating an average slope beneath the Butte of 7° E.

The outcrops of hard greenstone are hummocky and tree-covered. The eastern edge of the outcrop belt is irregular with "headlands" and "islands" of large green boulders or outcrops projecting from the main mass into the red lateritic soil.

The laterite forms a narrow zone of outcrop on the lower slopes of Jones Butte between the greenstone and the younger formations. Some of it has disintegrated to a red soil characterized by small, hard, hematitic pisolites. The remainder in this vicinity, and also farther south along the eastern edge of the greenstone, forms knolls of iron-rich, red to yellowish-red, pisolitic rock much like bauxite in appearance. Chemical analyses show, however, that it is a ferruginous laterite; this material has been described in State reports as a source of iron.

The best exposures of Edwin clay are north and south of the center of Jones Butte. The overlying soil has been stripped away, and pits have been opened in the hillside. Clay is only partly exposed along the west side of the hill on account of the overlap of the younger volcanic materials upon the laterite. East of the Butte the clay, laterite, and greenstone are covered by alluvium.

In several places the clay bed has been followed underground, and large amounts of material have been mined. The plan of the workings has been projected upon the map of the Butte (Pl. 1) and shows the extent of mining operations as of September 1941. Drilling on the east side of the hill reveals clay beneath the younger materials. The two tunnels projecting toward the center of the hill were driven through badly stained lateritic clay overlain in places by thin patches of good clay but more commonly capped by Ione sandstone or clay-bearing ash of the overlying volcanics.

Over most of the peripheral area the Edwin clay is 6 to 20 feet thick. Taken as a whole the formation strikes N. 18° E. and dips 7° E. in conformity with the slope of the underlying greenstone and laterite. Locally, however, the strike of the bed varies from N. 20 W. to N. 80 E., and the dip from 22° SE. to 1°–2° W. These local anomalies are due to compaction of the clay upon the irregular surface of the residual laterite.

The Edwin clay varies in color, texture, plasticity, and other physical characteristics, and two important divisions of genetic significance may be made. The lower half of the clay bed shows no stratification, and variations in color are gradational. On the other hand the upper part of the Edwin clay is stratified, and color differences between beds are sharp. Despite this contrast the upper portion has the same chemical, mineralogical, and refractory properties as the clay beneath it.

At the eastern extremity of the underground workings northeast of the center of the Butte, the Edwin clay is gray and more plastic due to the presence above it of two beds of brown lignite (Pl. 2). The upper bed is 4–5 feet thick and is separated from the 2-foot lower bed by a foot of gray, plastic clay. The lignite is stratified,

and the presence of many short roots and branches scattered throughout the underlying Edwin clay shows that lignite and clay are in part of the same age. The lignite may be correlated with beds of similar material formerly mined 2 miles farther north at Coal Mine No. 3. According to Piper *et al.* (1939) lignite is characteristic of the lower member of the Ione formation.

The anauxite-bearing sandstone of the Ione formation is disconformable upon the Edwin clay. The erosion interval is indicated by an irregular contact and fissures in the clay filled with sandstone. Throughout the Ione region and at Jones Butte the sandstone varies in coarseness, degree of cementation, and relative amounts of quartz, kaolinite, and anauxite. North and northeast of the center of Jones Butte the material is coarse and brownish white and carries large abundant quartz particles with white, pearly flakes of anauxite in a fine clay matrix. It is friable but when moist it forms a good hanging wall above the drifts and tunnels in the underground workings. In contrast, the material above the Edwin clay on the south side of the hill, best described as a sandy clay, is yellow to tan, fine-grained, and contains angular particles of quartz and minute plates of anauxite.

Between these two occurrences sand is not exposed. However, drill holes through it at intervals along the east side of the Butte show that the Ione sandstone here varies from brown to white, red, or blue and that it may be loose or compact, soft or indurated.

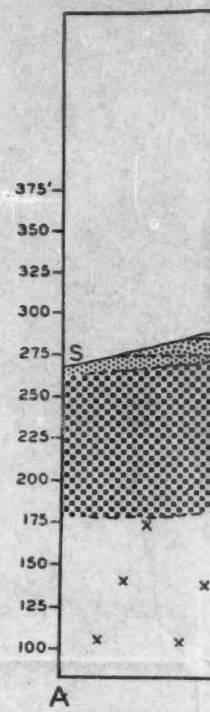
Despite such local variation the material may be regarded as a unit which has the same strike and dip as the underlying clay bed. The dip of 7° E. is primary upon the sloping surface of the underlying formation and is a local feature since the regional dip of the Ione sediments along the west flank of the Sierras is 2°–3° W.

Lying with distinct angular unconformity upon the sand of the Ione formation is the thick, horizontally bedded series of volcanic ash, tuff, and agglomerate which forms most of Jones Butte. It consists of two formations separated by a disconformity (Pl. 2).

The upper 146 feet of the series contains characteristically andesitic materials. Thick boulder beds in the lower part of the upper unit form a pronounced ledge traceable westward across neighboring hills of comparable height. Other members of this upper formation are sandstone, siltstone, and layers of andesitic breccia or agglomerate which presumably originated as mud flows. Piper *et al.* (1939, p. 61) have described these various members of the upper unit and have named the assemblage the Mehrton formation.

Beneath the horizontal cap of andesitic materials and above the eastward-dipping Ione sandstone is the Valley Springs formation, a wedge of flat-bedded rhyolitic detritus consisting of layers of ash, tuff, and ash agglomerate (Piper *et al.*, 1939, p. 71). The volcanic nature of the material distinguishes it from the sands and clays of the Ione formation, while the absence of andesitic fragments differentiates it from the Mehrton volcanics.

On the south side of the Butte quartz-anauxite sandstone almost identical to that of the underlying Ione formation is interbedded with ash agglomerate. Such an occurrence is described by Piper *et al.* (1939, p. 75) as common in the lower part of the Valley Springs formation. In the present case the quartz and anauxite not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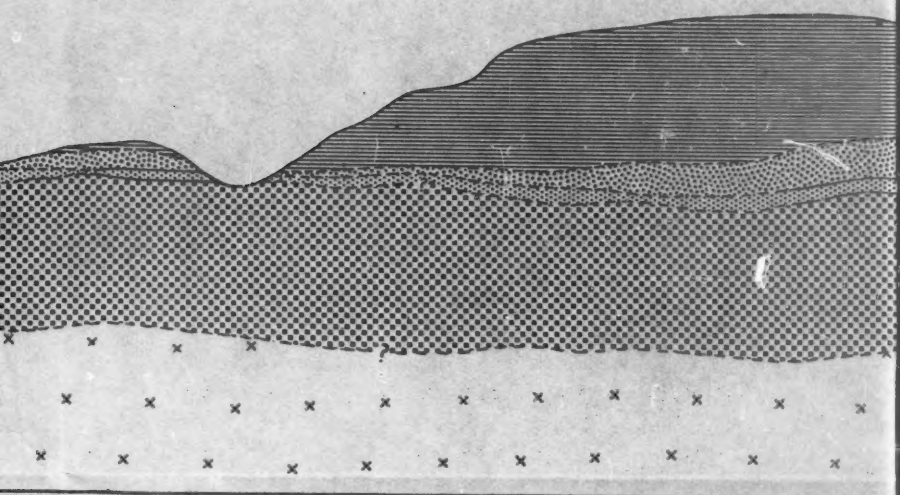

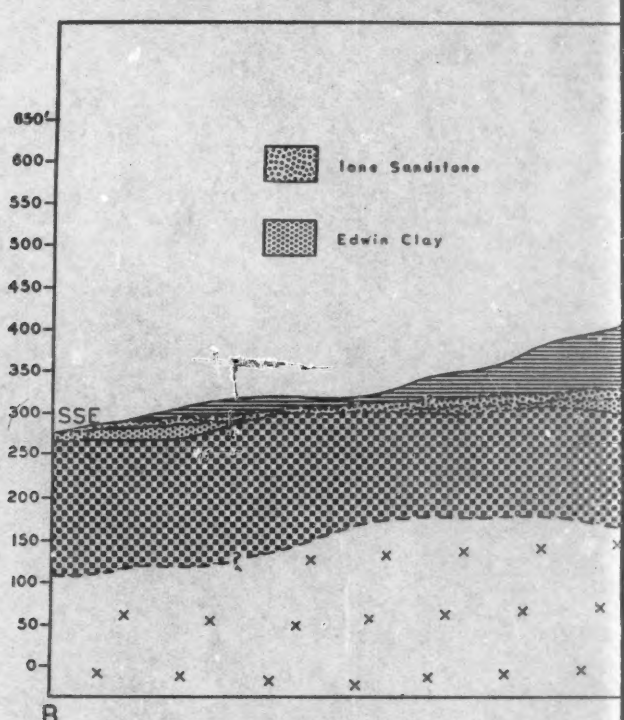
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Ione Sandstone
Edwin Clay
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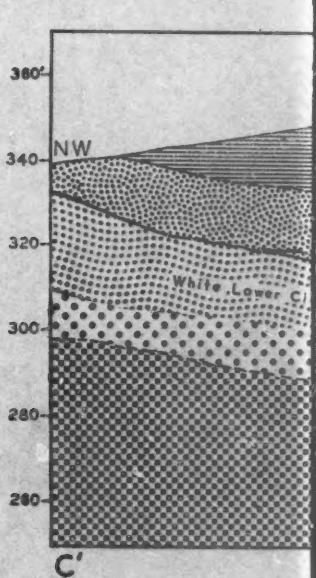
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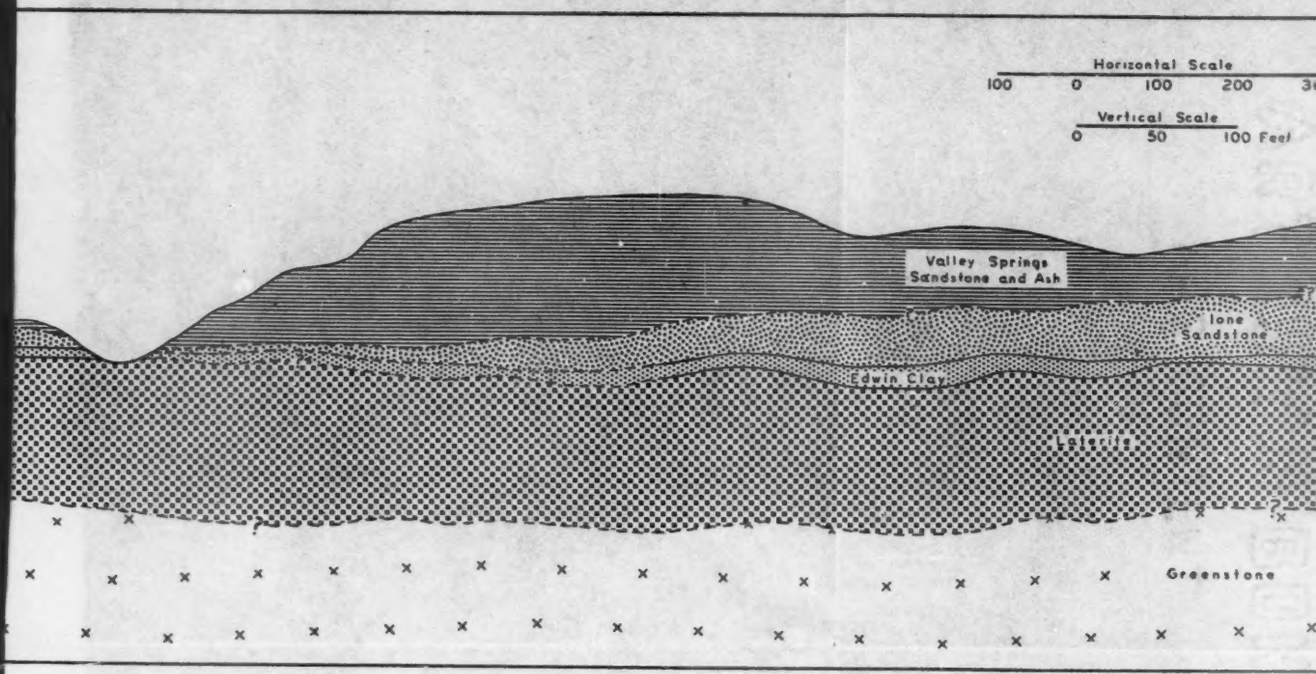
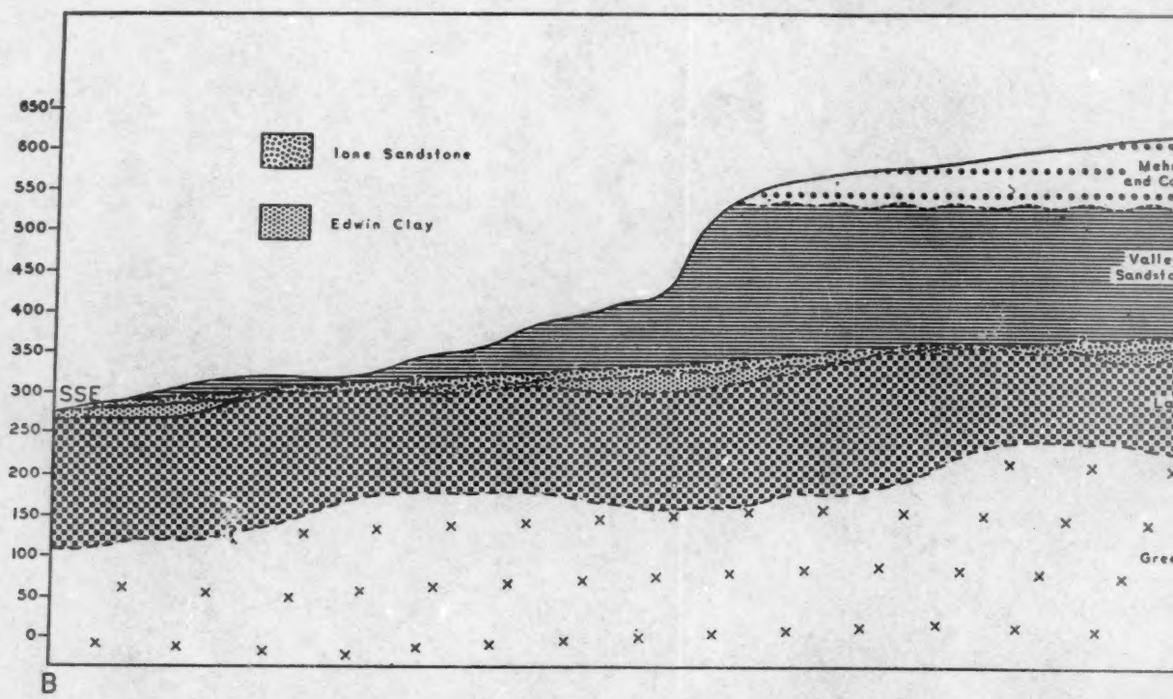


SECTION ALONG EAST SIDE OF JONES



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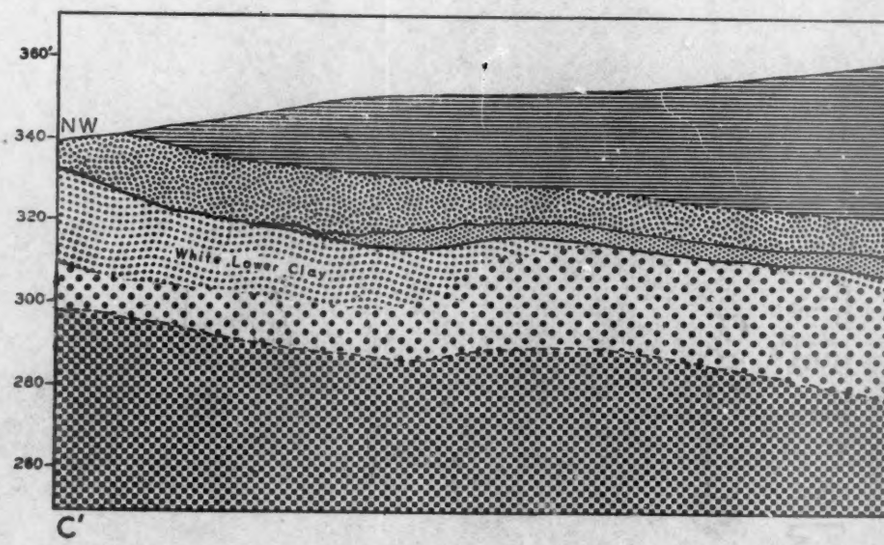

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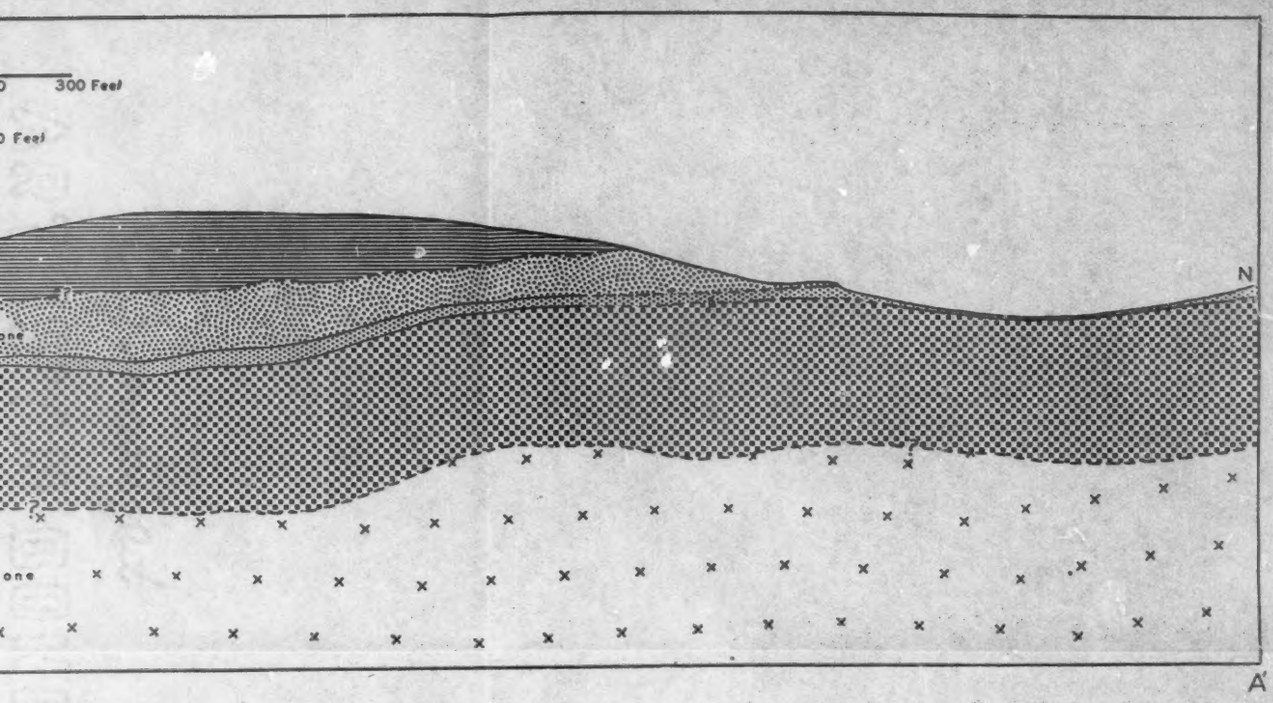
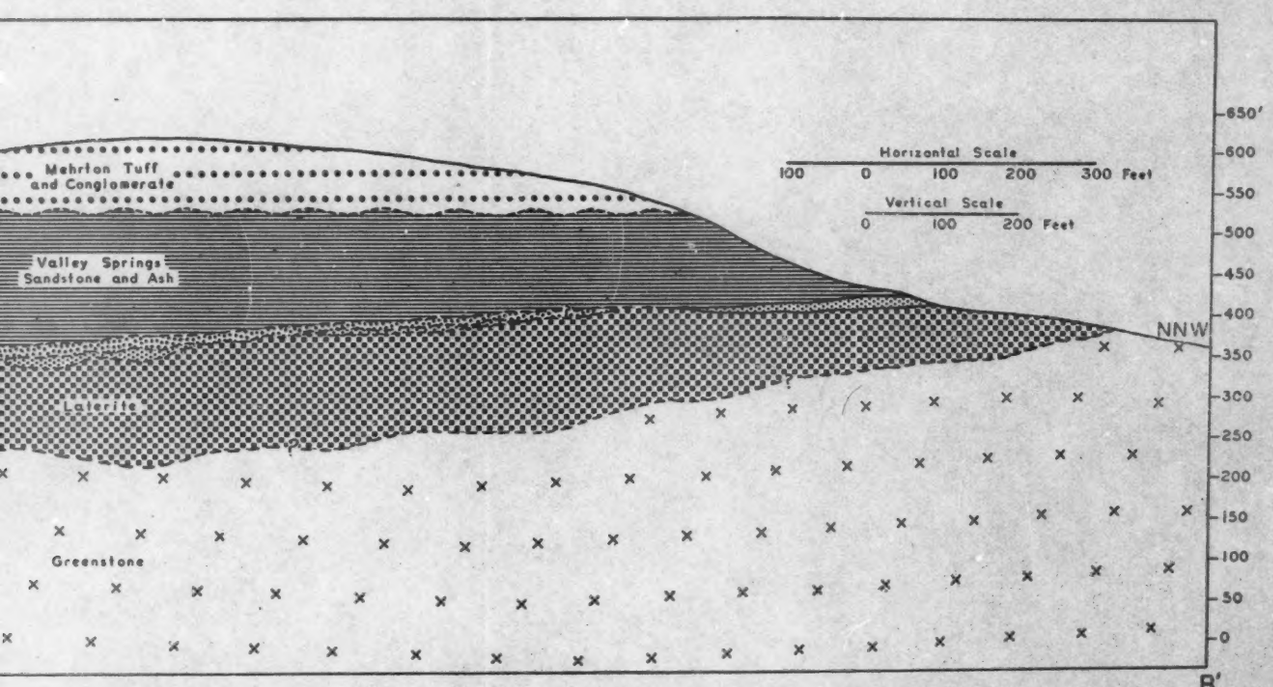
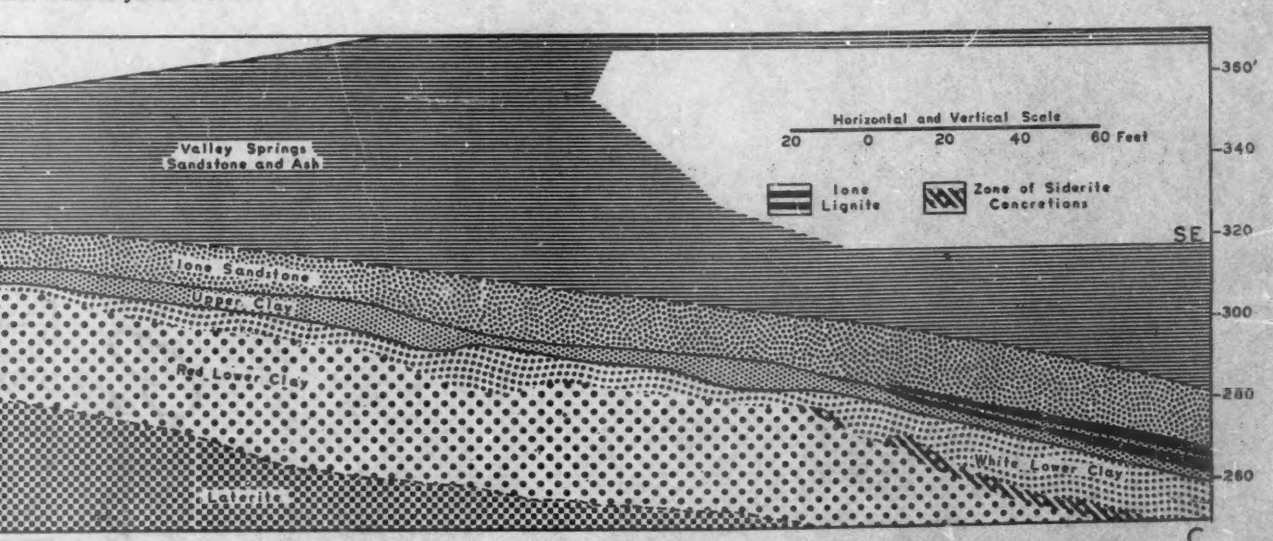


JONES BUTTE



TION ACROSS JONES BUTTE



SECTION, EDWIN NO. 2, JONES BUTTE

CROSS SECTIONS AT JONES BUTTE

only form a sandstone similar to that of the Ione formation, but these minerals are also mixed to a varying degree with volcanic detritus. This supports the contention of these authors that the sandstone beds "must be assumed to be the products of erosion and redeposition of materials from that [the Ione] formation."

Much of the lower part of the Valley Springs formation at Jones Butte is yellow, green, or bluish clayey ash. Turner (1894, p. 4) called this material "clay rock" and assigned it as the upper member of the Ione formation. At the Edwin No. 4 workings and in the two long tunnels driven toward the center of the Butte (*see* Plate 1) this "clay rock" is exposed above the Edwin clay where the Ione sandstone was removed by erosion prior to deposition of the volcanic materials. This relationship and the fact that the "clay rock" and Edwin clay have a similar appearance underground suggest derivation of the Edwin clay from the rhyolitic ash.

With this in mind, the relationship of the "clay rock" to the Edwin clay was investigated with some care particularly where the two formations are contiguous. The contact is not easily seen because of the similarity in color, but once observed it is sharp and irregular, indicating a disconformity. The ash is more plastic, moist, and porous than the Edwin clay. Near the contact it contains numerous hard, compact, angular lumps of Edwin clay which were broken out of the top of the clay bed and redeposited with the volcanic materials. According to Allen (1929, p. 415) the clay mineral in the ash itself is probably montmorillonite formed by alteration of pumice fragments.

Over most of the area these stratigraphic relations are not found since the rhyolite ash is separated from the Edwin clay by the Ione sandstone. Both the distribution and the disconformable relationships where clay and ash are adjacent indicate an absence of genetic relationship between the two formations. This fails to support the theory expressed by C. N. Schuette (1929) that the Edwin clay is an alteration product of the rhyolite ash.

GREENSTONE

INTRODUCTION

Greenstone is extensively exposed along the west side of the Sierra Nevada where it is associated with the Calaveras and Mariposa formations. Where it is well crystallized and granular this rock is considered intrusive, but it usually is amygdaloidal and has other characteristics of extrusive rocks.

The term "greenstone" is used for the metamorphosed volcanic rocks in this region. The various petrographic types were described by Lindgren (1894) and Turner (1894), who refer to the strictly igneous varieties as diabase and diabase porphyrite. They include in the "diabase series" small areas which consist more specifically of augite, hornblende, or quartz porphyrite. The clastic igneous materials are called fragmental diabase or diabase breccia. Actually other rock types may be distinguished, two of which are of importance to this problem—the andesite which forms the outlying ridge and the quartz diorite porphyry found in one place immediately east of the ridge. Both these rocks have been thoroughly altered to greenstones.

ANDESITE

The typical ridge-forming rock contains augite and plagioclase of intermediate composition as the only remaining primary minerals. Extensive porpylitization has obliterated all evidence of other original constituents.

The fresh rock is, in general, dark green, but in a few localities silicification has made it pale green to grayish white. The shade of green depends upon the abundance of small pyroxene phenocrysts and the amount of epidote and chlorite in the groundmass. Where the epidote is concentrated about the amygdules, the result is a mottling effect in shades of pistachio and dark green. Weathering of the rock produces a thin red to reddish-brown veneer of hematite.

The matrix is microcrystalline to cryptocrystalline and typically consists of secondary quartz, sericite, epidote, and chlorite. Occasionally tiny crystals of oligoclase are present in variable amounts.

The euhedral phenocrysts of augite and plagioclase are up to 2 millimeters long.

The augite occasionally shows evidence of alteration to chlorite, and on weathering hematite forms in the cracks and is sometimes retained as a boxwork structure. Usually, however, no apparent chemical alteration has taken place.

The characteristic plagioclase of the ridge-forming greenstone is oligoclase, but it ranges from calcic albite to sodic andesine. All the feldspar is altered although to varying degrees. Commonly it is partially or completely changed to sericite; elsewhere, it is covered by a brown dust of kaolin.

In the vicinity of Ione the greenstone is porphyritic, amygdaloidal, or fragmental. All three types may be found in the same rock, but the porphyritic and amygdaloidal varieties are most common.

Amygdaloidal textures characterize all the andesite but are more prominent in some places on the ridge than in others. The cavities form small circular or oval spots a few microns to a centimeter in diameter. Sometimes they have elliptical sections and are elongated until they become small, quartz-filled pipes 2 or 3 inches long and approximately half an inch wide. They are of importance because of their similarity to relict structures in the laterite. Fissures filled with hydrothermal minerals form streaks up to an inch across, and sometimes of considerable length.

Hydrothermal action has filled the cavities and fissures with characteristic minerals.

Idiomorphic crystals and flamboyant aggregates of quartz usually line the cavity or vein, frequently rimmed by a layer of chalcedony. Large, clear, interlocking plates of quartz not uncommonly fill the center of the cavity.

Chlorite is second to quartz in abundance and occurs in typical, relatively coarse, fibrous aggregates of green shreds with anomalous blue interference. It accompanies the anhedral, flamboyant vein quartz and is common in the amygdules.

Epidote is abundant in amygdules in distinct crystals and in velvety masses of tiny needles. Granular aggregates spot the matrix and sometimes form dark-green halos around the cavities, many of which are filled with clinozoisite. In all the andesite clinozoisite is found with epidote in cavities and veins, and to a lesser extent scattered through the matrix.

The centers of many amygdules are filled with calcite—apparently one of the latest minerals to form.

Small grains of pyrite, ilmenite, and less commonly of magnetite are often located with the quartz on the edge of the cavities but are also found in the matrix of the rock.

Grossularite is abundant in the cavities of one specimen. It usually lies between the quartz lining and the layer of chlorite. The crystals are occasionally roughly euhedral and always anomalous, showing weak gray birefringence.

One occurrence of a zeolite, heulandite, was noted in a large vein.

QUARTZ DIORITE PORPHYRY

On the northeast side of Jones Butte a drill struck the bedrock series at a depth of 90 feet. The rock encountered differs in several respects from the andesite: (1) the absence of amygdules; (2) the presence of chlorite pseudomorphs after pyroxene in place of fresh augite crystals; and (3) the presence of abundant euhedral, partly resorbed quartz phenocrysts. The feldspar is oligoclase like that in the ridge-forming greenstone. The rock has been altered hydrothermally but not so extensively as the andesite because of the absence of amygdules and scarcity of fissures. Nevertheless, epidote, clinozoisite, chlorite, sericite, and secondary quartz are common throughout the matrix.

Epidote and clinozoisite are found throughout the rock, often associated with chlorite in pseudomorphs after augite. None of the pyroxene remains, but sharp, eight-sided relics testify to its former presence. The sericite is largely restricted to replacement of feldspar phenocrysts although some of the feldspar, of apparently similar composition, contains only brown clay dust. The hydrothermal quartz is in irregular shapes which bear no resemblance to the much larger quartz crystals.

In addition to the hydrothermal minerals the matrix consists largely of potash and soda feldspars. Small apatite prisms are common, and grains of magnetite may be found.

The plagioclase phenocrysts are euhedral laths up to 2 millimeters in length. They have much the same appearance and composition as those in the andesite.

The large quartz crystals are the most striking feature of the rock. The euhedral phenocrysts are up to a centimeter in length and show the effects of having been resorbed before the matrix of the rock had solidified. The crystals show chains of bubbles and other inclusions which will be described in more detail later. A few of the phenocrysts have been partly replaced by chlorite and epidote (Pl. 5, fig. 1).

The rock corresponds to Turner's description (1894, p. 4) of quartz porphyrite which "appears in some cases to be intrusive into the diabase; in other cases, as in portions of the Gopher Ridge, there is a gradation between them." The number of drill holes at Jones Butte is insufficient to determine the nature of the change between the andesite and quartz diorite porphyry there, and the latter was not found on the surface in this vicinity.

LATERITE

PREVIOUS STUDIES

The ferruginous laterite in the vicinity of Ione has long attracted attention. Whitney (1865) and Turner (1894) write of it as an iron ore. Hill (1923, p. 27) calls the material laterite and gives three analyses furnished by the Stockton Fire Brick Company. Harrassowitz (1926, p. 487) mentions this occurrence along with other ferruginous laterite and bauxite of the North American Tertiary. Logan (1927, p. 135) refers to the material as impure hematite; while Dietrich (1928, p. 53) speaks of the "generally red, lateritic iron" which forms the floor beneath the clay

at Jones Butte. Schuette (1929, p. 197) also calls the material laterite but suggests the following origin:

"In forming the underlying laterite magmatic waters again were probably instrumental, and deposition of iron in the overlying tuff and agglomerate was probably an accompanying reaction during lateritization of the lower formation."

This concept of magmatic waters as an agent of lateritization is not in accord with the general concept of this process, nor does the evidence of this study support the theory that the solutions at Jones Butte were magmatic.

Allen (1929, p. 382) believed that the laterite near Ione was formed at the surface under tropical weathering conditions and that it exhibits part of the weathering profile typical of lateritization. As he pointed out, such a profile

"consists of four parts:

1. "Iron crust 1 to 8 feet thick forms the surface.
2. "Zone of concretion, 3 to 70 feet thick, made up chiefly of a porous mass of aluminum-iron hydrates, in which oolitic or pisolitic structure is extensively developed.
3. "Bleached zone, 15 to 80 feet in thickness, is composed of white or light-colored hydrous aluminum silicates. The name lithomarge has been applied to this part of the profile. Its most notable feature is the retention of the structure of the original rock, even where marked changes in chemical composition have taken place. Inclusions, quartz veins, and often details of texture are preserved.
4. "Fresh rock is reached by gradual transitions."

LATERITIC PROFILE

The complete profile is rarely seen in this region. Over most of the area where relations may be observed, only the clays of the bleached zone are present. Thus, an the east side of the Ione trough the Mariposa slates have been altered to highly argillaceous, red-stained material. This residual ferruginous clay is found at the western edge of the foothills, occasionally farther west where the slates project above the enclosing Ione sediments, or in drill holes. However, red stain derived from this "argillaceous laterite" extends westward for as much as half a mile from the edge of the slates and colors part of the clay, gravel, and sand of the Ione formation.

On the west side of the Ione trough, however, the andesite has been weathered to materials characteristic of the "bleached zone" and to laterite representing the "zone of concretion." Above the quartz diorite porphyry only the clay of the bleached zone is present in the single section available. In the formation of the Edwin clay deposit at Jones Butte these weathering products are important, and consequently a somewhat more detailed description of them is necessary.

MATERIALS OF THE BLEACHED ZONE

Two drill holes near the Jones Butte deposit of Edwin clay penetrated residual clays overlying the greenstone. Drill hole No. 10 was sunk in lateritic material at a point 300 feet northwest of the collar of the Edwin No. 2 shaft (Pl. 1). It reached a depth of 80 feet and penetrated to within a few feet of the andesite. Drill hole No. 9 penetrated the quartz diorite porphyry east of the andesite. It reached a depth of 90 feet, and its position is 1100 feet northeast of the collar of the Edwin No. 2 shaft.

In hole No. 10 the bottom 20 feet consists of coarse aggregates of kaolinite penetrated by a network of dark veins. Usually the veins consist of hematite, but in

places brown aggregates of radiating shreds suggest limonite pseudomorphous after chlorite. Flamboyant vein quartz, typical of the andesite, is common. The clay itself is birefringent and often forms short vermicular structures.

In the material above the quartz diorite porphyry in drill hole No. 9, hematite is much less common than in the alteration product of the more basic rock to the west. The quartz consists of large clear fragments containing chains and bands of fluid inclusions identical with those in the large phenocrysts of the parent rock. The clay itself has no vermicular structures but is birefringent and fibrous. X-ray-diffraction patterns show no difference in the two clay types but reveal that quartz is much more abundant in the material derived from the diorite porphyry.

In both these drill holes the residual clays pass upward into sedimentary materials derived from other sources, and consequently no evidence is forthcoming as to the nature of the transition from the bleached zone to that of the zone of concretion.

LATERITE OF THE ZONE OF CONCRETION

Pisolitic laterite is closely associated with the andesite and is limited to a narrow zone of outcrop along the east side of the ridge or to isolated pockets in the greenstone. The laterite of this type is resistant and forms low knolls strewn with irregular rough laterite boulders. Elsewhere, erosion has smoothed off the hillocks and exposed the red and black rock.

The pisolitic concentrations of hematite are typically closely packed and range from a sixteenth to half an inch in diameter. Usually the matrix is dark-red, argillaceous hematite which, however, is lighter-colored than the more concentrated materials. It is slightly less resistant than that forming the pisolites, so that the latter project as round knobs. Sometimes the matrix is tan and contains stringers of bluish-white, pearly masses of clay and gibbsite.

Microscopic study reveals the presence of quartz fragments derived from the amygdules in the andesite. Amygdular structures and some of the chlorite associated with the quartz in the parent rock have been preserved. The rims of the amygdules in the laterite are lined with gibbsite, whereas in the andesite the lining is commonly chalcedony. Figures 3 and 4 in Plate 5 show the similarity of structures found in the laterite to amygdules seen in the andesite. In such cases the existence of a zone of hydrous aluminum silicates between the andesite and the laterite is doubtful. Some of the matrix in the laterite is brown and isotropic and resembles cliachite.

Although much of the laterite closely resembles bauxite and contains appreciable amounts of gibbsite (Pl. 5, fig. 2), it is too ferruginous to be an ore of aluminum. This is shown by the following analysis obtained through the courtesy of the Stockton Fire Brick Company:[2]

	Per cent		*Per cent*
SiO_2	6.45	Na_2O	Nil
Al_2O_3	39.85	TiO_2	1.53
Fe_2O_3	31.35	Ignition loss	20.46
CaO	Nil		
MgO	0.36	Total	100.00
K_2O	Nil		

[2] Analysis by Curtis and Tompkins, Ltd., San Francisco.

EDWIN CLAY

GENERAL DESCRIPTION

The clay at Jones Butte is of two readily distinguished types separated by a disconformity. Since the deposits vary from 10 to 20 feet in thickness each type is well represented.

The lower clay is white at the top but changes gradually to a dark red at the base of the section. The bottom of the deposit is nowhere exposed. Bedding is recognized only occasionally under the microscope, while megascopically the clay is characterized by an irregular fracture which reveals a network of many interlacing cylindrical structures up to a quarter of an inch in diameter.

The upper clay possesses the same mineralogic and refractory qualities as the white clay of the lower type, but its beds are sharply marked by local disconformities and differences in color. It has conchoidal fracture but not the cylindrical textures characteristic of the lower clay. The upper clay is overlain disconformably by the Ione sandstone.

LOWER CLAY

Megascopic description.—The lower clay represents a perfect gradation from red, iron-stained material at the base to pure, white clay at the top (Pl. 3, fig. 1). The gradual change of color has been effected by the leaching action of organic solutions which have removed the hematite stain from the upper part of this type but have not been able to bleach the entire deposit. The leached half is of commercial importance, as the white clay at the top is extremely refractory.

The color of the red clay is broken only by fine, anastomosing lighter veinlets which outline circular, rectangular, or elliptical sections (Pl. 4, fig. 1). Stringers of siderite run down the center of the veinlets and sometimes are concentrated into round specks. The iron carbonate is present wherever the surrounding clay has been slightly bleached. The clay itself is hard, massive, and breaks conchoidally since the veins are not sufficiently large to affect the type of fracture.

Siderite is most abundant just above the base of the section where it is found in concretions which range from a fraction of an inch to over a foot in diameter (Pl. 2). The carbonate disappears as the clay becomes lighter in color.

The hematitic clay grades upward into material in which the veins of white clay are broader and the areas of unleached hematite between them are cylindrical in shape with more nearly round cross sections than in the underlying clay (Pl. 4, figs. 2, 3). The result is the formation of vertical or near vertical structures, more resistant than the surrounding bleached clay, which project from the surface as rounded ridges usually about a quarter of an inch wide and up to 3 inches in length, and bear a marked resemblance to plant stems (Pl. 3, fig. 1, C). The red hematitic core and the gradation into the incipient structures found in the red clay at the base of the section indicate that these cylinders resulted from leaching and not from replacement of organic matter.

Upward the clay changes from red to brown to white (Pl. 4, figs. 4, 5, 6) and becomes softer and less compact although the cylindrical masses of more ferruginous

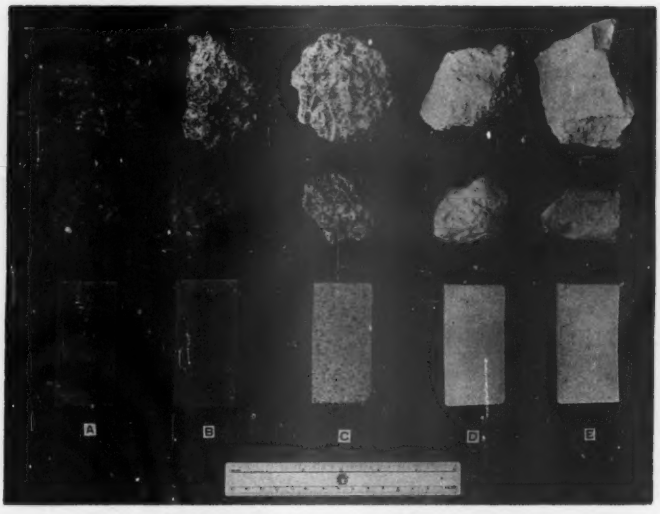


FIGURE 1. FIVE VARIETIES OF EDWIN CLAY
Showing raw sample, fired lump, and briquette. A–D, lower type; E, upper type clay.
A. Red, hematitic clay from base of exposed part of lower clay zone.
B. Mottled, ferruginous clay above *A* from which some of the hematite has been leached.
C. Mottled, brown and white clay above *B* showing cylindrical structures formed by percolating organic solutions.
D. Highly refractory, white Edwin clay found above *C* at the top of the lower clay zone.
E. Upper Edwin clay which lies disconformably upon varieties *A–D*.

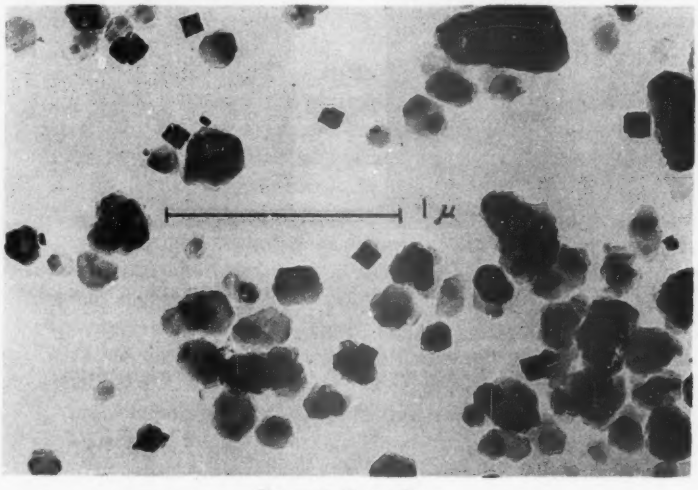


FIGURE 2. EDWIN CLAY
Electron micrograph showing aggregates and hexagonal-shaped plates of kaolinite. Square shapes suggest pseudo-cubic crystals of anatase.
(Courtesy of RCA laboratories)

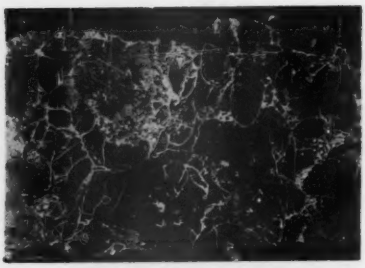


FIGURE 1

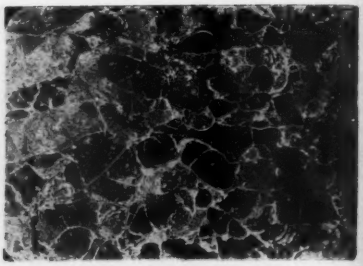


FIGURE 2

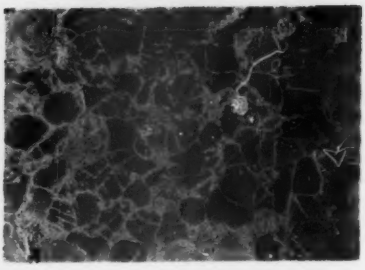


FIGURE 3

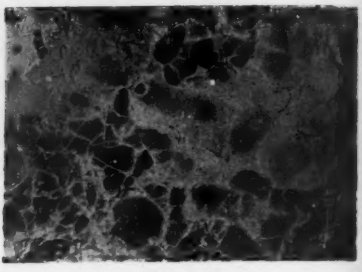


FIGURE 4

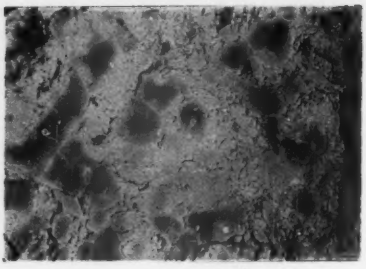


FIGURE 5

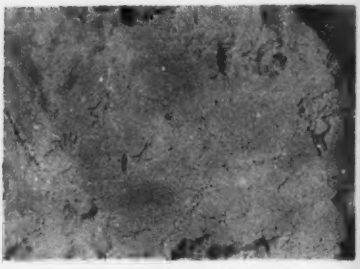


FIGURE 6

LOWER EDWIN CLAY SERIES SHOWING SECTIONS OF THE STRUCTURES CAUSED BY LEACHING. (FULL SIZE)

(1) Slightly leached hematitic clay at the base.
(2–5) Intermediate stages in the transition.
(6) Refractory clay near the top with only a trace of the original staining.

material are more resistant than the broadened veins of white clay. Even where all trace of brown or pink has been removed from the cylindrical masses by continued leaching, the structures are still visible because they are slightly more resistant than the matrix. These features in the white clay indicate that the hematite content here was originally as high as that in the unleached red clay at the base of the series. As a result of these structures the fractured surface of the lower clay type is always rough in contrast to the smooth fracture of the upper clay. Also because of these areas of slightly greater resistance, the lower clay is more friable than other varieties. Both the friability and the texture, however, are not so apparent near the base of this zone where the material is hard and compact or near the top where it becomes more plastic.

Throughout the lower clay are scattered dark-red pisolites up to an eighth of an inch in diameter. They are similar to the extremely hard pisolites in the laterite and have resisted the leaching.

Microscopic description.—The refractive index of the clay aggregates is 1.576 ± .005, and the birefringence is .002. The index is high for kaolinite, possibly because the clay is filled with specks less than 1 micron in size, which have an index of refraction greater than 1.500. Allen believed the clay to be halloysite and explained the high index ("1.570 ± .005") by reference to investigations by Splichal (1922) who "has shown that the index of refraction increases to 1.57 as the water decreases" (Allen, 1929, p. 382). Allen also states that the clay is isotropic, but the clays studied by the author all show weak birefringence in strong incident light and consist of fibrous aggregates which show positive elongation and preferred orientation, as though the clay particles were roughly aligned during deposition.

In the clays there are also distinctive opaque, irregular to euhedral hexagonal crystals averaging 10 microns in size and anisotropic prismatic crystals (10 by 3 microns, average) with optical properties suggesting apatite.

The electron micrograph of the white lower clay (Pl. 3, fig. 2) indicates that the clay is extremely finely crystalline. The micrograph and an electron-diffraction pattern were taken at the RCA laboratories on material supplied by the author. Individual crystals average 0.1 micron in diameter and are infrequent because of the marked tendency to form aggregates. However, the distinct single crystals show hexagonal outlines typical of kaolinite and are similar to published electron micrographs of kaolinite (Alexander, *et al.*, 1943; Shaw and Humbert, 1941).

Small cubic shapes may also be observed in the micrograph. Many show a steplike structure at the corners suggesting parallel growth. Such a feature is characteristic of anatase (TiO_2) which is tetragonal but pseudocubic and is an authigenic mineral in clays. Its occurrence here would explain the abnormally high titanium content and the high refractive index of the aggregates. Probably these crystals are the minute specks visible in thin sections.

Larger mineral fragments, ranging from 0.6 mm. to 0.015 mm., are common. Quartz, zircon, magnetite, ilmenite, and pyrite are abundant; tourmaline, hornblende, and hematite grains are common; and andalusite is present. Masses of gibbsite, which on a smaller scale resemble the pisolites in the laterite (Pl. 5, fig. 2), are seen in the red clay at the base but disappear near the middle of the lower clay

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zone. However, the presence of aluminum hydroxide has been traced throughout the Edwin clay series. Pisolites are found throughout the lower clay, and some of them contain aggregates of tiny gibbsite crystals. The hematite particles which give the clay its deep red color at the base of the series are too small to be seen in thin section. These minerals and their significance will be discussed in more detail later.

The red iron stain was removed from the clay along the paths of the entering solutions and was precipitated as siderite (Pl. 5, fig. 5). The carbonate also replaces grains of quartz (Pl. 5, fig. 6), and some of it in turn is altered to limonite and hematite.

Evidence of bedding in the lower Edwin clay is rare, but occasionally a bedding plane is defined by a string of mineral grains lying along an irregular crack through the section. One such crack contained fragments of ilmenite, zircon, tourmaline, quartz, pyrite, and siderite. The clay on one side does not differ from that on the other.

X-ray-diffraction data.—Table 1 lists the interplanar spacings obtained from X-ray-diffraction patterns of the Edwin clay. The photographs were taken by Professor Paul F. Kerr who also furnished the Brooklyn kaolinite and Liege halloysite for purposes of comparison. The diffraction pattern of the Edwin clay resembles the kaolinite more than the halloysite.

This resemblance is also brought out by relative sharpness of the lines in the diffraction patterns of the three materials. Figure 3 is a photometer record of a similar portion of three of the films obtained by Professor Kerr. It was taken on a Leeds and Northrop microphotometer kindly made available by Dr. Wheeler P. Davey. The steep slope of the curves is the result of change in intensity of the background. The halloysite lines are much more diffuse than those of the kaolinite. The lines from the Edwin clay, although not as sharp as those of the Brooklyn kaolinite, are much more distinct than those of the halloysite.

Other X-ray-diffraction patterns taken by the author show that the iron stain at the base of the lower clay series is hematite and not one of the hydrous iron oxides or hydroxides. Quartz, siderite, and the other minerals are not abundant enough to contribute to the lines on the patterns of Edwin clay.

Electron-diffraction data.—An electron-diffraction pattern was made at the Princeton laboratories of the Radio Corporation of America from a sample of white lower Edwin clay. According to Hillier, "measurements on the diffraction rings agree quite accurately with those obtained from kaolinite." No electron-diffraction data on halloysite could be obtained.

Differential thermal analysis.—As a further check on the character of the Edwin clay several thermal analyses were made with the apparatus of the Department of Ceramics of the Pennsylvania State College. The results are shown in Figure 4.

Ferruginous clay from the base of the lower Edwin clay zone was analyzed. The resulting curve shows the typical reactions for kaolinite with an endothermal peak at 600° C. and a strong exothermal peak at 950° C. The absence of other endothermic peaks in the thermal analysis of this clay (which by chemical analysis contains 12.18 per cent Fe_2O_3) substantiates X-ray evidence that the red iron stain is due to hematite rather than to some iron hydroxide.

TABLE 1.—*Interplanar spacings from X-ray-diffraction patterns*

Kaolinite, Brooklyn		Upper Edwin clay		Lower Edwin clay		Halloysite, Liege, Belgium	
d	I/I_1	d	I/I_1	d	I/I_1	d	I/I_1
8.020	0.10	7.944	0.10				
7.818	0.20						
7.280	0.50	7.308	0.20	7.258	0.10	7.192	0.50
7.011	0.50	7.022	0.20	7.018	0.50		
4.806	0.05	4.781	0.05	4.861	0.20	4.777	0.30
4.365	0.20	4.370	0.30	4.373	0.70	4.378	1.00
4.111	0.30	4.133	0.05				
3.966	0.10						
3.886	0.10	3.892	0.10	3.880	0.30		
3.802	0.10						
3.694	0.05						
3.537	1.00	3.527	0.90	3.527	0.90	3.533	0.05
3.488	0.10			3.480	0.10		
3.319*	0.10						
3.076	0.05						
2.805	0.10	2.809	0.01	2.825	0.10	2.810	0.20
2.743	0.10	2.718	0.01	2.732	0.10		
2.547	0.30	2.534	0.40	2.538	0.40	2.541	0.10
2.507	0.20						
2.478	0.30	2.473	0.30	2.472	0.40		
		2.416	0.05				
2.338	0.20	2.362	0.10	2.361	0.30		
2.325	0.80	2.317	0.60	2.316	0.50	2.306	0.40
2.276	0.60	2.277	0.10	2.263	0.10		
2.237	0.05			2.213	0.10		
2.170	0.10	2.179	0.05	2.159	0.10	2.192	0.03
2.117*	0.03						
2.054	0.02			2.076	0.05		
1.977	0.50	1.967	0.10	1.968	0.20		
1.929	0.30	1.930	0.05	1.928	0.05		
1.883	0.20	1.878	0.20	1.880	0.30		
1.825	0.40	1.827	0.05	1.826	0.05	1.843	0.05
1.775	0.30	1.762	0.30	1.777	0.20	1.757	0.06
				1.732	0.01		
		1.690	0.10	1.690	0.10		
		1.673	0.05	1.673	0.10		
1.659	0.60	1.655	0.30	1.657	0.30	1.665	0.20
1.631	0.10	1.629	0.20	1.628	0.20	1.619	0.20
1.608	0.40			1.598	0.05		
1.573	0.20						
1.528	0.50			1.531	0.05		
1.4804	0.80	1.477	1.00	1.478	1.00	1.471	0.90
1.441	0.20	1.447	0.20	1.451	0.30		
1.421	0.10			1.425	0.01		
1.408	0.05			1.409	0.01		
1.383	0.10			1.387	0.01		
1.366	0.08			1.371	0.01		
1.351	0.08	1.355	0.10	1.355	0.10		
1.335	0.20	1.332	0.05	1.334	0.05		
1.298	0.40	1.306	0.02	1.303	0.03		
1.278	0.40	1.278	0.40	1.279	0.30	1.275	0.30
1.260	0.20	1.259	0.10	1.260	0.10		
1.242	0.30	1.243	0.05	1.246	0.05		
1.229	0.30	1.229	0.20	1.228	0.30	1.229	0.30
1.196	0.10			1.205	0.01		
1.187	0.10	1.187	0.05	1.186	0.01		
1.161	0.03	1.160	0.05	1.162	0.01		
1.121	0.05			1.138	0.01		
1.106	0.03			1.109	0.05		
1.089	0.01						
1.076	0.01						
1.047	0.01			1.049	0.05		
1.041	0.01			1.041	0.05		
1.018	0.01			1.020	0.08		
1.010	0.01						

*Quartz lines.

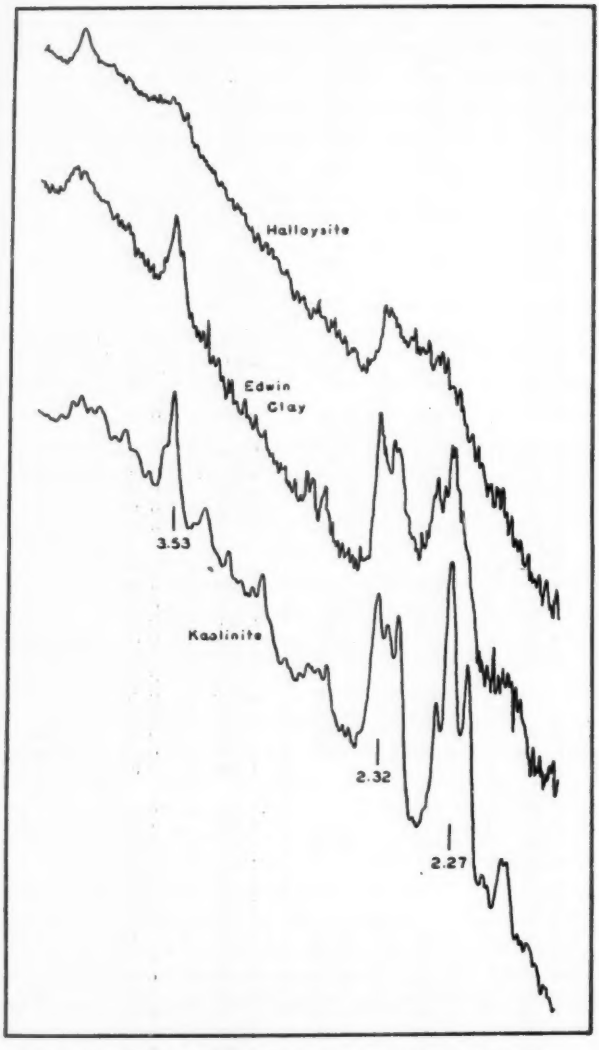

FIGURE 3.—*Photometer record of X-ray-diffraction patterns*
Figures refer to interplanar spacings.

In the white clay from the top of the lower Edwin clay zone the kaolinite effects are supplemented by a small endothermic reaction between 290° and 350° C. This is in the range of the endothermic heat effect of gibbsite as shown by Norton (1939) and Grim and Rowland (1942; 1944).

Refractory qualities of the clay.—Data obtained from firing tests show striking

evidence of the gradation from ferruginous material, sample No. 9, at the base to the white clay, sample No. 15, at the top of the lower clay series (Table 2; Pl. 3, fig. 1).

Chemical composition.—Analyses 2 and 3 of Table 3 show the decrease in iron content from the iron-stained clay at the base of the lower clay zone to the typical white Edwin clay. The theoretical composition of kaolinite is given for comparison.

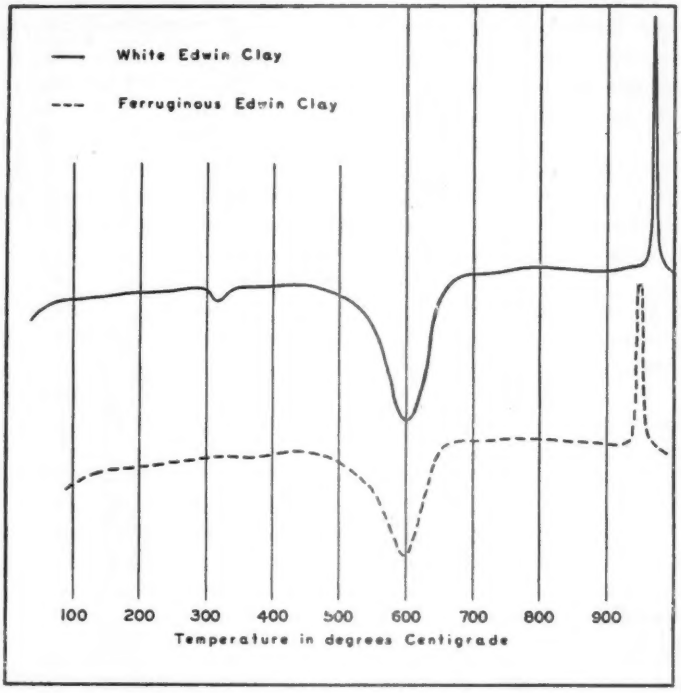

FIGURE 4.—*Differential thermal analysis curves*

These analyses provide additional data for determining the composition of the Edwin clay and show that the clay mineral is kaolinite. Four analyses of anauxite and 10 of kaolinite (Ross and Kerr, 1931, p. 163) show that the SiO_2:Al_2O_3 ratio of anauxite ranges from 308:100 to 235:100, while that of kaolinite ranges from 205:100 to 185:100.

The Edwin clay gives SiO_2:Al_2O_3 ratios of 191:100 and 184:100 (Table 3). These ratios fall near those of the kaolinite listed by Ross and Kerr and rule out the possibility that the Edwin clay is anauxite.

However, although the ratios compare favorably with those of other kaolinite and help establish the identity of the clay mineral, they fall below the theoretical 2:1 ratio for kaolinite. In fact the excess of alumina over silica is slightly greater than that expressed by the ratios since small amounts of quartz are present in the

clay. This is of particular interest since two analyses of hand-picked kaolinite crystals from the Ione formation a mile south of Ione (Ross and Kerr, 1931, p. 163) give 196:100 and 200:100 for SiO_2:Al_2O_3.

TABLE 2.—*Pyrometric cone equivalents and linear shrinkage in Edwin clay*

Sample	P.C.E.	Approximate temperature	Linear shrinkage (%)
		°C.	*At cone 13, 1350°C.*
9	23–26	1580–1595	17.5
10	29	1640	16.5
11	31–32	1680–1700	18.0
12	33–34	1745–1760	20.0
13	plus 34	1765	20.5
14	plus 34	1765	20.75
15	plus 34	1765	21.75

TABLE 3.—*Chemical analyses*

	(1)	(2)	(3)	(4)	(5)
	%	%	%	%	%
SiO_2	46.50	35.31	41.32	48.10	52.75
Al_2O_3	39.50	31.44	38.14	32.60	32.23
Fe_2O_3		12.18	1.34	1.25	1.80
FeO		2.33			
CaO		trace	0.36	0.12	0.22
MgO		0.26	0.02	0.10	0.28
K_2O		0.03	0.39	0.34	1.87
Na_2O		0.24	0.11	0.14	
TiO_2		2.48	2.62	2.20	
H_2CO_3		0.53			
Combined H_2O	14.00	13.96			
Ignition Loss		1.18	15.69	15.15	11.38
Total.....	100.00	99.94	99.99	100.00	100.53

(1) Theoretical composition of kaolinite.

(2) Lower Edwin clay, sample No. 9; analysis by Booth, Garrett, & Blair, Philadephia, Pa. (*Note:* If recomputed without Fe_2O_3, FeO, and H_2CO_3 the similarity to analysis (3) is more apparent with SiO_2 41.59, Al_2O_3 37.03, etc.)

(3) Average of four analyses of Edwin clay by Curtis & Tompkins, Ltd., San Francisco. Courtesy of the Stockton Fire Brick Company.

(4) Cheney Hill clay; analysis by Curtis & Tompkins, Ltd.; courtesy Stockton Fire Brick Co.

(5) Ione clay (Eastern type); average of three analyses from the Gage, Fancher, and Harvey pits (Dietrich, 1928).

Gibbsite aggregates in the ferruginous clay at the base of the Edwin series have been noted. Thin sections of the other clays in the series do not show these aggregates although some hematitic pisolites contain gibbsite crystals. However, the presence of a hydrate of aluminum in veins, small pisolites, and tiny specks in all varieties of the clay was ascertained by use of morin dye (Feigl, 1935, p. 236; Tscheischwilli *et al.*, 1939).

It was found that if a methyl alcohol solution of the dye was poured over a dry lump of the specimen, veinlets and other concentrations of gibbsite gave a bright-green fluorescence, while the rest of the clay was not affected.

Thus it is evident that the relative excess of alumina shown by the chemical analyses of the Edwin clay is due, at least in part, to the presence of gibbsite and possibly other aluminum hydrates. Taking this into consideration, the chemical analyses support other evidence that the clay mineral is kaolinite.

UPPER CLAY

Megascopic description.—The clay in the upper half of the deposits is separated by a disconformity from the lower clay. In places, some of the material at the top of the lower zone has been removed, and in one area on the 400–450-foot levels of the Edwin No. 2 workings yellow, limonitic staining and fissures filled with bright-red, ferruginous clay indicate surface alteration.

The Edwin clay in the upper zone (Pl. 3, fig. 1) has been deposited in thin lenses and sheets on the surface of the lower type and shows distinct planes of stratification. None of the cylindrical textures are present, and the uniform mass fractures conchoidally. The most plastic varieties are coherent and, when moist, form a compact clay wall.

The lenses in the upper clay are white, buff, tan, chocolate, pink, and orange, and, although the shades may grade horizontally into each other over some distance, they appear uniform in a single exposure. Small, broadly spaced, red or brown pisolites, similar to those in the lower clay, occur in all the lenses.

Near the surface the clay sometimes is mottled yellow by limonite from surface waters. In the plan of the 100-foot level in the Edwin No. 2 workings the distribution of this type of clay follows an elongate winding pattern suggestive of a stream course. The presence, also, of vertical streaks of bright-red clay suggestive of fissures strengthens the possibility of near-surface alteration.

Where lignite overlies the section the clay of both zones is stained gray to black. It is commonly moist and very plastic and contains irregularly scattered, carbonized roots and branches. The carbon does not affect the commercial importance of the clay, but concentrations of pyrite are also common and lower the value of the material.

Microscopic description.—In thin section the upper clay is similar to the white clay of the lower zone. The fibrous character is more apparent due to slightly higher birefringence. The index of refraction of the aggregates is 1.585 ± .005. Minute specks are abundant, but the hexagonal-appearing opaque crystals are much less common than in the lower type. Prismatic crystals believed to be apatite are present in about the same abundance.

Of the larger fragments ilmenite is the most abundant. However, magnetite, quartz, and zircon are common. Tourmaline, hornblende, andalusite, pyrite, hematite, and occasional clusters of sericite are also present. The pisolites form sharp boundaries with the surrounding clay. They consist of hematite and occasionally are traversed by cracks containing fine crystals of gibbsite. Individual masses of this mineral were not observed.

No siderite is seen in the clay of the upper zone. However, particularly in some of the pink varieties, white veinlets of lighter clay anastomose through the material, suggesting that small amounts of the iron have been removed by the same process which operated in the more ferruginous clay of the lower zone.

TABLE 4.—*Refractory data on the Ione clays*

No. of deposit, deposit and sample	P.C.E.	Approx. temp.	Linear shrinkage
		°C.	%
Edwin Clay: Lower type			
1 Edwin #2, sample #13	plus 34	1765	20.50
sample #15	plus 34	1765	21.75
Edwin Clay: Upper type			
1 Edwin #2, buff to tan	plus 33	1750	19.25
orange to chocolate	plus 34	1765	20.00
white to buff	34	1760	21.25
gray	34	1760	18.75
mottled yellow	32–33	1722	18.75–19.25
Edwin #10, typical white clay	34	1760	21.00
Cheney Hill Clay			
2 Smoky Corners, quartz rich	33–34	1752	14.00
quartz poor	34	1760	16.00
3 Cheney Hill, lower clay bed	plus 33	1750	9.00
upper clay bed	33–34	1752	8.50
upper, minus quartz	plus 34	1765	21.25
4 Woolford Pit, lower clay bed	plus 33	1750	9.50
upper clay bed	31–32	1690	12.75
5 China Gulch, typical clay	33	1745	8.75
6 Chalk Hill, some quartz*	33	1745	15.75
quartz rich	33	1745	7.50
7 Fancher Pit, lower clay bed*	33–34	1752	11.75
upper clay bed	plus 33	1750	11.25
Eastern Ione Clays			
8 Clark pit, typical clay	minus 33	1740	18.75
9 Airplane Pit, 5th bed from top	32–33	1722	14.25
2nd bed from top	32–33	1722	16.00
10 Chitwood Pit	30–31	1665	13.75
11 Wallen Pit	31–32	1690	6.50
12 Bacon Red Pit, red clay	23	1590	14.25
yellow clay	27–28	1610	10.25
purple clay	18	1490	13.75
13 Ione Fire Brick Pits	30–31	1665	9.50
Bacon Red type	29–30	1645	9.50

* Data pertains to Cheney Hill type clay beds although bulk of the clay in these deposits is of the Eastern Ione clay type.

X-ray-diffraction data.—Diffraction patterns taken by the author and by Professor Kerr of the clay of the upper zone are similar to those of the kaolinite below (Table 1). Other minerals are not abundant enough to be represented by lines on the films

Refractory qualities.—In general the clay of the upper zone is not quite so refractory as that at the top of the lower zone. Tests were made at the laboratories of the Stockton Fire Brick Company on samples taken from the Edwin No. 2 workings (Table 4). Unless iron-stained, the clay fuses at 1750°–1765° C., while its linear shrinkage at 1350° C. varies from 18.75 to 21.25 per cent.

PROPERTIES OF OTHER IONE CLAYS

GENERAL DESCRIPTION

Exposures of Ione clay are found from Merced Falls to Oroville, California. Over this distance (150 miles) the clays show many local variations (Allen, 1929). In the vicinity of Ione the Ione clays vary in color, texture, amount and size of quartz grains, and refractory qualities. For purposes of comparison with the Edwin clay a number of other Ione clays were studied in detail. The deposits selected for this comparison fall into two groups, depending upon their position in the area of Ione deposition: (1) clays from deposits near the western edge of the area—*i.e.*, farthest from the source, and (2) deposits in the middle and on the eastern side of the area (Fig. 2).

CHENEY HILL CLAY

The clays on the west side of the Ione trough are designated the Cheney Hill type from their characteristic appearance at deposit No. 3 (Fig. 2). They are all similar particularly in that: (1) The clay is well stratified, in beds 3 to 10 feet thick. It is massive, compact, and commonly brittle and typically shows conchoidal fracture. (2) In deposits nearest the greenstone ridge the clay at the base of the section has been stained red to a varying degree. (3) Angular quartz grains are abundant. They average 0.5 mm. in length but range from 0.05 to 1.5 mm. The abundance of the grains distinguishes the Cheney Hill clay from the relatively quartz free Edwin clay, while their large size differentiates this type from many of the other quartzose Ione clays in which the grains average less than 0.05 mm. (4) The clay is exceedingly fine and shows the fibrous character typical of the Edwin clay. It resembles the upper rather than the lower type in that the tiny, hexagonal-shaped, opaque crystals are not common. The index of refraction of the clay aggregates is $1.582 \pm .005$. (5) Zircon, tourmaline, hornblende, andalusite, magnetite, ilmenite, and pyrite occur in relatively the same abundance in each deposit. The amounts of hematite and gibbsite vary. (6) X-ray-diffraction patterns of the different clays are identical and, provided the quartz has been removed, are the same as those of the Edwin clay. (7) The fusion points of the nonferruginous clays is between 1745° and 1760° C., while the linear shrinkage at 1350° C. varies from 8.5 to 16.25 per cent depending on the amount of quartz (Table 4). It is of significance that washing the quartz from one sample raised its fusion point from cone 33–34 (ca. 1752°) to plus 34 (ca. 1765°) and its linear shrinkage at cone 13 (1350° C.) from 8.5 to 21.25 per cent. This corresponds to the refractory qualities of the best Edwin clay. (8) The chemical composition of the Cheney Hill variety is given in Table 3. If it could be assumed that all the alumina in the analysis is present in the clay mineral alone, recalculation on the basis of the theoretical composition of kaolinite would reveal an excess of approximately 10 per cent SiO_2. However, since morin tests show that an aluminum hydroxide is present in appreciable amount, the excess of SiO_2 above that required for kaolinite must be greater than 10 per cent. Nevertheless this can easily be accounted for by the abundant grains of quartz.

Except for the large amounts of angular quartz, the clays near the western edge

of the area of Ione sediments are alike and notably similar to the Edwin clay. The deposit at Smoky Corners (deposit No. 2, Fig. 2) is transitional between the quartz-rich and quartz-poor varieties. The clay on the east side of the deposit is quartzose with a linear shrinkage of 14.00 per cent at cone 13. Quartz grains diminish in abundance (but not in size) toward the west until within approximately 100 yards the westernmost clay contains only one-tenth the amount of grit. The linear shrinkage increases to 16.25 per cent. The clay here is locally called the Edwin type although it is slightly inferior to that found at Jones Butte.

EASTERN IONE CLAY

The Ione clays from the middle and eastern parts of the area of Ione deposition are similar in themselves but notably different from the Cheney Hill and Edwin varieties: (1) The clays are well stratified in beds 2 to 10 feet thick although in a few places, due to local uniformity of material and deposition or to masking of the bedding by iron stain, they are 25 to 30 feet thick. A great variety of colors and textures are represented, in a few places being like those of the clay farther west. (2) In deposits nearest the Mariposa formation on the east some of the clay and overlying sand is stained red or mottled red and yellow by iron oxides produced by lateritization of the adjacent bedrock. (3) Near the middle of the area most quartz grains do not exceed 0.05 mm. The average fragment is from 0.01 to 0.005 mm. long. Although very abundant throughout the clay, they are easily passed over because of their small size. Toward the east these fragments increase in size up to 0.1 mm. in diameter. At the same time much larger grains, averaging 0.5 mm., appear in the clay and increase in abundance toward the east. (4) These clays are not so fine-grained as the Cheney Hill clays, individual particles being visible at magnifications of 1000 times. They are birefringent and also in fibrous aggregates which, however, have an entirely random orientation. The clay samples from deposits near the middle of the area of Ione sedimentation give the index of $1.572 \pm .005$ for the clay aggregates.

Minute specks, possibly anatase, are abundant. The clays are also characterized by hundreds of tiny needles, the largest being about 0.015 mm. long and very thin. Many resemble the rutile inclusions seen in quartz. Others, less abundant, occur in slightly larger and stouter prisms (ca. 0.025 by 0.007 mm.) and resemble apatite. (5) Larger crystals of zircon are abundant; apatite, ilmenite, pyrite, magnetite, and hematite are common; and tourmaline, hornblende, and andalusite are present. The morin test reveals the presence of a hydrate of aluminum in variable amount in these clays. In general, samples from the middle of the area of Ione deposition (*e.g.*, deposits 8 and 9, Fig. 2) either give no reaction or show weak fluorescence in veins or specks. On the other hand, samples from deposits 11 and 12 on the eastern edge of the area show many veins and pisolites which fluoresce brilliantly. (6) X-ray-diffraction patterns are the same as those of Cheney Hill clays except that quartz lines are prominent even for the clays which contain none of the larger crystals. (7) The P.C.E. and linear-shrinkage values for some of the Eastern Ione clays are given in Table 4. The first three pits listed are farther west than the last three. The pyrometric cone equivalents of this group of Ione clays vary from 30 to 33

(1650°–1745° C.) indicating that they fuse at lower temperatures than do the Cheney Hill and Edwin types. The linear-shrinkage values, ranging from 6.5 to 18.75 per cent, indicate large amounts of quartz commonly present. (8) The chemical composition of the typical Eastern Ione clay is given in Table 3. The large amount of SiO_2 is probably the result of the abundant quartz shown by microscopic and X-ray examination.

ORIGIN OF THE EDWIN CLAY

INTRODUCTION

Allen (1929) and Piper *et al.* (1939) have shown that the Ione sediments formed delta deposits in the Eocene sea in the area just west of the Sierra Nevada. A long period of stability characterized by a tropical climate and lateritic weathering was ended prior to Ione time by uplift and erosion. The highly refractory clays of the Edwin and Cheney Hill types were derived from the upper, most thoroughly decayed part of the weathering profile. The mixture of extremely fine-grained clay with coarse-grained quartz and heavy minerals confirms this and shows that little sorting took place during transportation.

Allen (1929, p. 404) first suggested that the clay at Jones Butte possesses especially good refractory qualities because of derivation from the upper part of the lateritic profile, but no one has explained in detail the unusual characteristics which set apart the Cheney Hill and Edwin varieties from the coarser, less refractory types to the east. It is hoped that the following data will provide a satisfactory explanation.

The Edwin clay is a quartz-poor variety of the Cheney Hill type; at Jones Butte it has been thoroughly exploited. For this reason the following discussion centers about this deposit, which is also of greatest geological and mineralogical interest.

The problem of origin has been dealt with from the standpoint of (1) source of the materials; (2) mixing of these materials to form the deposit; and (3) leaching of some of these materials.

SOURCE OF THE MATERIALS

General statement.—Although the deposit at Jones Butte is largely fine-grained, relatively pure Edwin clay, there are associated with it minerals which came from the east with the clay and materials derived from a source to the west. Some of the latter are important in that they give to the deposit the qualities which make it of particular interest. The source on the east was the same thoroughly weathered Sierra bedrock which furnished the clay. In addition it yielded quartz of several types, zircon, tourmaline, hornblende, and andalusite together with some other minerals which are not diagnostic. The source to the west was the andesite of the adjacent ridge and the ferruginous laterite which had formed from it in pre-Ione time. This source furnished fine hematite in the form of a staining pigment, gibbsite in crystal aggregates, pisolites containing hematite and/or gibbsite, and slightly abraded fragments of coarse clay typical of the first stage of greenstone weathering. None of the heavy minerals listed above were found in the andesite. Quartz,

feldspar, and other materials of the quartz diorite porphyry, which lies beneath and slightly to the east, were not seen in the clay of the deposit.

Heavy minerals.—The heavy minerals in the Edwin clay are zircon, tourmaline, hornblende, andalusite, magnetite, ilmenite, and pyrite. Andalusite grains may be as long as 0.1 to 0.2 mm., while the others rarely exceed 0.06 mm. in length. All the minerals are found in the Cheney Hill varieties and in the other Ione clays to the east. Like the quartz, they are more abundant in these other types than in the Edwin clay for reasons which will be made apparent later.

Allen (1929, p. 375) found that ilmenite, zircon, tourmaline, and andalusite are common in Ione sands, while magnetite and hornblende are present. As he points out, "the presence of andalusite is noteworthy, for it has been described by Turner as a contact mineral, produced by the granitic intrusions into the Mariposa slates." Zircon, tourmaline, and hornblende all occur in the rocks of the Sierras, while magnetite, ilmenite, and pyrite occur in both the Sierras and in the outlying andesite. This suite of minerals, intimately mixed with the clay, indicates that the source of most of the kaolinite of the Edwin clay deposit at Jones Butte was the bedrock of the Eocene "Sierra Nevada."

Quartz types.—Inclusions in the quartz grains found in the clay are extremely variegated. In about 750 grains from the Cheney Hill and Edwin clays studied in detail there were recognized: 11 minerals; 2 varieties of minute, acicular needles; and 8 well-defined varieties dependent upon the shape and distribution of fluid inclusions. Study of 100 grains in the andesite and quartz diorite porphyry and their weathering products led to the recognition of a number of other quite different characteristics among the inclusions in the quartz. Although there is a great deal of overlapping of varieties of inclusions three basic groups can be recognized. Each of these can be subdivided further into three types to which inclusions in quartz were first assigned by Mackie (1899) in his classic paper:

I. *Regular* inclusions, in which he noted "the following minerals—quartz as an inclusion in quartz, chlorite, muscovite, biotite, rutile, apatite, zircon, garnet, magnetite, titaniferous iron."

II. *Acicular* inclusions which "appear as fine dark lines running in all directions through the quartz grains."

III. *Irregular* inclusions which "are almost always fluid lacunae with or without gas bubbles."

In relating these types to the source material Mackie found that

"it may be stated as a fairly general law that acicular and irregular inclusions preeminently abound in the quartz of granite; that the regular group is to be found in various proportions, but always in relatively larger numbers in the quartz of gneiss and the younger schistose rocks."

Where types overlap, acicular take precedence over regular which in turn take precedence over the irregular inclusions.

The quartz from the Edwin and Cheney Hill clays is distinguished from that in the andesite and the quartz diorite porphyry by the inclusions, which are classed in three basic groups. These are summarized in Table 5.

The summary shows that the typical quartz of the Cheney Hill and Edwin clays did not come from the andesite or quartz diorite porphyry. The only other likely source was the rock of the Sierra bedrock series to the east which furnished the heavy minerals. Unfortunately the significance of the quartz inclusions is not fully estab-

TABLE 5.—*Summary of quartz inclusions*

Group No.	Occurrence	Description of quartz	Inclusions	Description of inclusions
1	Quartz of andesite and weathered products	Hydrothermal—flamboyant and interlocking masses of vein quartz, clear crystals in cavities (often twinned in short spindle-shaped laths).	*Acicular:* very rare*	
			Regular: chlorite, sericite, epidote, hematite.	Replace quartz along edges or occur in isolated crystals.
			Irregular: either few or very abundant large bubbles.	
2	Quartz of quartz diorite porphyry and weathered products	Primary—large, partly resorbed phenocrysts. Large and small irregular masses in the matrix.	*Acicular:* very rare*	
			Regular: apatite, zircon, hornblende, monazite, in the phenocrysts. Chlorite in quartz of the matrix.	Euhedral to subhedral, rarely longer than 30 microns. Chlorite is along thin, parallel cracks giving streaked effect.
			Irregular: oriented chains and bands of bubbles on a clear background in the phenocrysts (Pl. 5, fig. 1)*	Chains commonly outline roughly hexagonal sections.*
			A few larger, round, pink bubbles.	Bubbles are from 10 microns to less than 0.2 micron in diameter.
3	Quartz of Cheney Hill and Edwin clay	Angular fragments averaging 0.5 mm. in diameter.	*Acicular:* abundant*	
			(1) thick needles	Red-brown, up to 100 microns long. Sometimes striated oblique to edges. Called rutile by Mackie (1899), Tyler (1936).
			(2) extremely fine needles	Up to 202 microns long. Average 0.7 micron in width.
			(3) Coarse, intermediate, and/or fine needles in varying abundance.	Typical grain may contain four or five of medium size.
			Regular: apatite, hornblende, tourmaline, zircon, muscovite, hematite, sericite.**	
			Irregular: scattered or grouped in chains, single layered sheets, or broad zones.	Clear grains rare.* Bubbles range from 8 microns to less than 0.2 micron, often contain minute bubbles within them. Chains usually parallel to subparallel.

* Outstanding characteristic of group.

** Also an unidentified mineral, possibly a member of the chabazite group: indices *ca.* 1.48, pink, rhombohedral crystals which appear almost cubic (angle of 85° 30′), extinction symmetrical to rhomb outlines, birefringence of one crystal 0.014.

lished since time did not permit a study of the quartz in the rocks of the Sierra Nevada. However, a quantitative study based on Mackie's three types shows that in 295 grains from the Cheney Hill and Edwin clays 146, or roughly 50 per cent, are irregular (containing fluid inclusions only); 60, or 20 per cent, are regular (with mineral inclusions predominating); and 89, or 30 per cent, are acicular (with needles predominating). Such a distribution is, according to Mackie (1899, p. 152), indicative of the quartz in granite. It is likely in the present instance that the source of the quartz was not a single type of parent rock, but probably most of the quartz originated in the granite and granodiorite of the Sierras.

The concept that the heavy minerals and the large quartz grains were derived from the Sierras is important. The Cheney Hill and Edwin clays in which they are found are the most fine-grained clays in the region, lie farthest from the eastern source, and are separated from the source rocks by coarser Ione clays which have abundant, but much finer-grained quartz. Thus the streams and currents carrying the Cheney Hill clays westward must have been strong enough to transport the heavy minerals and the coarse-grained quartz. Since the transporting agents afforded no means of sorting, these clays must have originally been finer-grained than the other varieties of Ione clay. If this reasoning is correct, it substantiates Allen's belief that the clay at Jones Butte—the Edwin type—was derived from the upper, most thoroughly leached part of the weathering profile. The same holds true for the more common Cheney Hill clay of which the Edwin is merely a quartz-poor variety.

Materials from a western source.—The deposits of Cheney Hill clay adjacent to the outlying andesite ridge have been mixed with products of weathering of the basic rock. These materials came from various parts of the lateritic weathering profile that had formed from the andesite in the tropical climate of pre-Ione time. Weathering processes destroyed the few primary minerals in the rock and most of the secondary, hydrothermal minerals, although quartz and a little chlorite occur in some of the laterite (Pl. 5, fig. 4). At the same time new minerals, chiefly kaolinite, hematite, and gibbsite, were developed. At the Jones Butte deposit of Edwin clay these materials are mixed with the kaolinite, heavy minerals, and quartz from the Sierras.

QUARTZ: The identification of hydrothermal quartz derived from the andesite to the west is doubtful. In the Edwin clay a few grains show twinning characteristic of the secondary quartz in the andesite. Since this feature was not seen in any of the quartz from the Sierras this might be considered diagnostic. However, together with a few notably clear grains with chains of inclusions, it is the only evidence of quartz derived from the andesite.

HEMATITE: More diagnostic, and more important to the history of the Edwin clay deposit, is the hematite. In the Ione clays on the eastern edge of the Ione trough, as well as in the Cheney Hill varieties on the western side, the clays next to the bedrock have been stained red by hematitic pigment derived from the adjacent laterite. At the Jones Butte deposit the hematite originally stained all the lower Edwin clay, although farther east the stain dies out.

Although locally the staining produces mottling much of the stained clay at Jones Butte is deep, solid red. All varieties show the typical fine grain, heavy

minerals, and quartz types characteristic of the Cheney Hill clays. Individual particles of hematite are not visible, but X-ray-diffraction patterns show that the red stain is caused by this mineral.

GIBBSITE: This mineral is common in veins and masses in the lateritic material surrou. ding the Butte. Near the base of the lower Edwin clay it occurs in crystalline aggregates of low birefringence. They commonly show the irregular fractures (somewhat resembling shatter cracks) and other features typical of similar structures in the laterite (Pl. 5, fig. 2).

In the other clays of the Edwin series at Jones Butte, and also in the other Cheney Hill clay deposits, an aluminum hydrate is present in veinlets and disseminated throughout the kaolinite.

PISOLITES: Pisolites containing hematite and/or gibbsite are common in both the lower and upper Edwin clays. Some of the upper, stratified clays are, in fact, characterized by the presence of scattered red pisolites $\frac{1}{16}$ to $\frac{1}{8}$ inch in diameter. Some of these are argillaceous, while others resemble the hard pisolites found in the laterite.

CLAY: The first stage of bedrock alteration formed a zone of clay which in places still lies upon the andesite and quartz diorite porphyry. Small, rounded masses of this material are found in the Edwin deposit. The two clay types are easily distinguished since that formed from the greenstone is much coarser, more birefringent, and typically occurs in short vermicular aggregates. X-ray patterns show that the material is kaolinite.

The presence of these materials in the Edwin clay shows that the deposit at Jones Butte does not consist entirely of minerals derived from an eastern source. The uplift which caused erosion and removal of the weathered Sierra rocks also affected the outlying andesite ridge. The most obvious result was the removal of the hematite which stained the clay being washed in from the east. But coarser fragments from the laterite, and even from the clay zone beneath it, were eroded and transported a short distance down the slopes of the cove to be mixed with the fine clays.

FORMATION OF THE EDWIN CLAY DEPOSIT

The Edwin clay deposit at Jones Butte differs from the other deposits of Cheney Hill clay in that: (1) The quartz and heavy minerals are notably less abundant in the Edwin clay, and (2) some of the materials from the laterite to the west are much more abundant in the Edwin clay.

Both these differences are due to special conditions. Jones Butte lies in a "cove" or embayment in the andesite (Fig. 2). As long as this rock formed a ridge, even of low relief, materials washed into the shallow Eocene sea in the Ione trough could enter the cove only from the east. From this direction the Sierra kaolinite, quartz, and heavy minerals came.

It is logical to assume that the waters of this bay were relatively quiet permitting only the clay and limited numbers of heavy minerals and large quartz grains to enter. On the other hand, the lateritic weathering products of the andesite formed three sides of the cove and with erosion and slight transportation were concentrated in the waters at the foot of the slopes.

Evidence supporting both these assumptions is found by comparing a stratigraphic

section taken in the Edwin No. 2 workings within the embayment with a section from a drill hole 1000 feet farther east on the eastern edge of the cove (Fig. 5).

Section B has at the base the quartz diorite porphyry found east of the andesite. The inclusions in the large quartz phenocrysts belong to Group 2 (Table 5). No diagnostic heavy minerals are found in this rock.

The diorite porphyry is capped by 10 feet of blue, gritty clay (bed No. 2 in the diagram) largely derived from it. Quartz is common and has the same inclusions as that of the rock below. The matrix is streaked with limonite and consists of sericite, kaolinite, and fine angular quartz fragments often strung out in definite directions giving the material a schistose appearance and indicating that it has been reworked. Abundant crystals of zircon found in the heavy-mineral fraction indicate an influx of material from the east.

Bed No. 3 consists of 10 feet of brown and red sandy clay also derived from the reworked diorite porphyry. In addition a grain of microcline and one of andalusite suggest an outside source.

In the middle of the section is 44 feet of clay of the Cheney Hill variety. The inclusions in the abundant quartz grains are all of Group 3, the typical heavy minerals are present, and the clay is the fine-grained, fibrous, weakly birefringent Cheney Hill type. The lower 20 feet is the bluish-white, quartzose clay so common in the Cheney Hill deposits. Bed No. 5 above it is the same except that it is mottled red and bluish-white. The red represents the influx of hematite derived from the ridge to the west. As the clay passes westward into the lower Edwin clay it becomes solid red, gibbsite and larger pisolites are more prominent, and all except a few grains of quartz disappear. Similarly the 4 feet of clay in bed No. 6 is a quartz-rich variety of the upper Edwin clay. Quartz fragments are smaller and less abundant than in the clay below, but they are much more common here than in the continuation of this bed in section A and other places within the embayment. The remaining 10 feet of clay in bed No. 7 contains more abundant, larger quartz fragments and in this respect is more typical of the Cheney Hill group than No. 6. However, the clay itself, although largely of the fine, Cheney Hill type, has been adulterated with a little coarser clay material. While it may be considered a coarse phase of the Cheney Hill clay, it indicates that the source material changed with continued erosion. A quartz-poor variety of this phase is present at the top of the upper clay zone in most of the Edwin workings although not in section A. It contains more quartz than the typical upper Edwin clay and represents the beginning of an influx of coarser materials terminating with the deposition of the Ione sandstone.

The succeeding beds of lignite, and the typical anauxite sandstone at the top of the column aid in correlating the two sections. The quartz in the sandstone has the same inclusion types as does that in the Cheney Hill clays.

In summary, these and other sections yield the following data pertaining to the formation of the Edwin clay deposit:

(1) The Edwin clay passes eastward into quartzose clay identical with that seen in other deposits of the Cheney Hill type. The transition from quartz-rich to quartz-poor types takes place within 1000 feet. Employees of the Stockton Fire

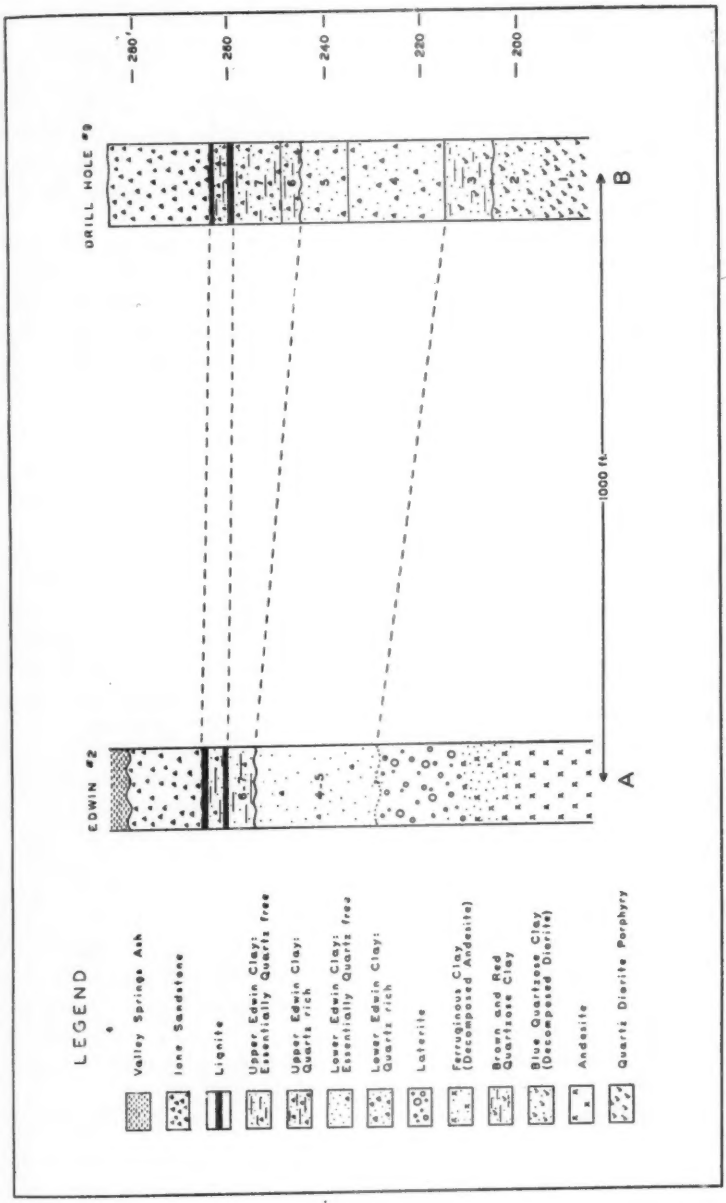

FIGURE 5.—*Columnar sections at Jones Butte*

Brick Company state that in some of the old workings, now inaccessible, the northeast limit of clay mining was determined by the sudden appearance of quartzose clay.

(2) The lower type of Edwin clay within the embayment (beds Nos. 4, 5, section A) was entirely stained by fine hematite. Pisolitic aggregates of gibbsite are common in the lower part of the series but disappear rapidly toward the top where the hydrate of aluminum is more disseminated throughout the clay. Except for local concentrations, pisolites of hematite are evenly distributed throughout the series. All these materials decrease in abundance to the east.

This information suggests the following steps in the history of the formation of the Edwin clay deposit: During deposition of the lower Edwin clay the adjacent ridge was low, and eroding agents were weak, since the fine hematite and gibbsite were the chief materials supplied from this source. However, over a long period 20 to 30 feet of fine, massive, poorly bedded Edwin clay was deposited. Once within the waters of the bay the fine clay, hematite, and gibbsite particles settled quietly and formed a thick, essentially homogeneous deposit.

The conditions required for the deposition of the lower Edwin clay were finally interrupted. After a thorough leaching of the hematite some of the bleached clay was affected by surface alteration, some of it was slightly reworked, and in one or two places erosion removed it and exposed the underlying laterite.

The succeeding influxes of clay were more spasmodic giving rise to the lenses characteristic of the upper type (beds Nos. 6, 7 in the sections). Pisolites are abundant, and fragments of the residual clay found above the greenstone are present. On the other hand, the upper clay shows no evidence of having been thoroughly stained, as was the lower clay, probably because either (1) in the erosion interval between deposition of the lower and upper clays the fine hematite was removed from the weathered laterite leaving available only the coarser pisolites; or (2) fine hematite was furnished as before but was reduced by organic acids in the waters of the cove. As will be shown in the following section, such solutions were introduced in the interval between deposition of the lower and upper clays and may well have been present during deposition of the upper Edwin clay. The fact that the larger, hard pisolites of hematite are evenly distributed and of the same size in both lower and upper clays suggests that the character of the source did not change.

A maximum of 10 to 15 feet of the upper clay was deposited. At the top of this zone clays richer in quartz suggest that currents were washing coarser material into the bay. This is confirmed by the lignitic materials conformably above the Edwin clay on the eastern side of the deposit. Carbonized branches and roots of all sizes are found and increase in abundance and size to the east. Even well within the cove at the eastern edge of the Edwin No. 2 workings a log over 3 feet wide is exposed for more than 15 feet in the roof of one of the drifts.

Following another period of erosion, the coarse Ione sand was deposited as a blanket over the clays and the coal.

LEACHING OF THE DEPOSIT

Were this the complete history, the Edwin clay deposit would be of much less commercial importance since during deposition the lower clay, which represents

two-thirds of the deposit, was so badly stained with hematite as to make it valueless as a refractory. However, waters containing organic acids and CO_2 percolated through the ferruginous clay prior to the deposition of the upper stratified clay and leached the hematite from the upper half of this lower zone. Some of the iron was precipitated farther down as siderite, and some of it was carried off to the east.

This phase of the history of the deposit has been worked out in detail by a study of the siderite in the clays at the Butte (Fig. 5). No siderite is found in the quartz diorite porphyry at the base of section B, but in the residual gritty clay in bed No. 2 iron carbonate is concentrated in rosettes up to 3 mm. in diameter. On reworking of this material fragments containing siderite, clay, and quartz were redeposited in the succeeding layers of brown and red gritty clay of bed No. 3. No new siderite was added. Thus, solutions acted on the weathered bedrock prior to the deposition of the Cheney Hill clays, either during or following the period of lateritic weathering.

In the lower 20 feet of Cheney Hill clay (bed No. 4 in section B), siderite replaces the quartz fragments. In the hematitic clay of bed No. 5 no siderite is found, and the hematite was not leached. To the west, however, in the lower clay of section A, the organic solutions had their maximum effect. The light color and refractory qualities of the clay in the upper half of this zone were completely restored by the removal of the hematite. The base of the zone was only slightly leached but received much of the precipitated siderite.

This, the second influx of solutions, followed the deposition of the lower Edwin clay. The waters were active only within the embayment since the hematite stain in bed No. 5 of section B was not affected. Siderite was precipitated throughout the lowest part of the clay in section A, and particularly in the lower levels of the Edwin No. 2 workings where concretions 1 or 2 feet in diameter may be found (Pl. 2). In addition some of it was carried farther eastward in solution until it concentrated around or replaced grains of quartz as in bed No. 4 of section B.

The upper Edwin clay in sections within the embayment contains no siderite. However, pink varieties sometimes show veinlets of lighter clay as if slight amounts of iron had been removed; and in the Edwin No. 3 deposit slightly darker shades in the light-colored clay have the shape of pisolites, as if leaching had been effective here. In section B, bed No. 6, which correlates with the upper clay farther west, has no siderite, but the clay above it and below the lignite resembles bed No. 4 in that siderite surrounds and replaces the quartz fragments.

This indicates that there was a third influx of carbonate solutions following deposition of the upper Edwin clay and before the lignite was washed in.

Thus, at Jones Butte siderite was formed three times. In section B it is found in beds 2, 4, and 7. Lignite occurs above bed No. 7, while above each of the beds is a disconformity. The lignite and the lower two disconformities can be correlated with three lignite seams described in the section of Coal Mine No. 3, a mile north of Jones Butte (Irelan, 1888, p. 111). Probably these seams extended as far south as Jones Butte, but only the uppermost escaped subsequent erosion.

There is evidence therefore of a direct connection between the siderite and the presence of lignite. The formation of the iron carbonate also is related to the leaching of hematite from the lower Edwin clay. Stagnant water and plant decay,

attendant upon the formation of lignite, produced not only CO_2 but also solutions capable of dissolving hematite. Organic acids produced under such conditions are capable of dissolving ferric oxide (Harder, 1919; Gruner, 1922; Harrar, 1929). At Jones Butte such acids apparently dissolved large amounts of iron, while the CO_2 present in the same waters precipitated much of it as siderite.

SUMMARY AND CONCLUSIONS

The sediments of the Ione formation in the vicinity of Ione, California, lie in a shallow, elongate trough bounded on the east by the foothills of the Sierra Nevada and on the west by a discontinuous, low ridge of outlying greenstone of the Sierra bedrock series. The Ione clays within this area may be separated into two groups: Deposits near the western edge of the trough consist of exceedingly fine-grained clay with abundant quartz fragments averaging 0.5 mm. in size—the Cheney Hill clay. In contrast, in the middle and eastern parts of the trough coarser clay contains many quartz fragments averaging only 0.01 to 0.005 mm. in size. This clay—the Eastern Ione clay—is less refractory than the other and shows greater linear shrinkage on firing.

The Edwin clay is a variety of the fine-grained Cheney Hill type. It is essentially quartz free, is the most refractory of the Ione clays, and has the highest linear shrinkage.

The Edwin clay is found in a large deposit 3 miles northwest of Ione beneath a hill known as Jones Butte. The clay in the lower two-thirds of the deposit grades from white at the top to red clay at the bottom. Megascopically it shows no stratification but is characterized by an irregular fracture resulting from breakage around many interlacing cylindrical tubes which are slightly more resistant than the matrix. The change from white to hematitic clay is gradual. Near the top of the series the red color first appears in the cylindrical structures, while farther down the matrix also becomes more and more ferruginous. The presence of the structures in the white clay at the top of the lower clay zone indicates that the entire zone was originally stained with hematite. Leaching at the top of this zone was effected by atmospheric waters charged with organic acids. The lower Edwin clay contains quartz, various heavy minerals, tiny hematite particles, aggregates and veinlets of gibbsite, pisolites containing hematite and gibbsite, and pisolites of coarse clay.

The clay in the upper third of the deposit is well stratified and shows a smooth, conchoidal fracture rather than the cylindrical textures characteristic of the lower clay. Although the beds vary in color there is no vertical gradation into clay stained by hematite or into the lower clay. Gibbsite (or some similar aluminum hydrate), quartz, heavy minerals, and pisolites are present in the same abundance in the upper as in the lower clay.

Data collected from microscope studies, X-ray- and electron-diffraction patterns, firing tests, differential thermal analyses, chemical analyses, and electron-microscope photos show that the clay in the upper part of the deposit is similar to that below it. In addition these studies indicate that the clay mineral is kaolinite.

To ascertain the origin of the Edwin clay three aspects of the history of the deposit

at Jones Butte were studied: (1) the source of the materials; (2) the formation of the Edwin clay deposit; and (3) the leaching of the deposit.

The materials in the clay were derived from two sources. The hematite, gibbsite, and pisolites are all found in the lateritic weathering products of the andesite ridge. The heavy minerals (zircon, tourmaline, hornblende, andalusite, magnetite, ilmenite, and pyrite) are all in rocks of the Sierra Nevada. Inclusions in the quartz grains indicate an eastern source. The fact that the heavy minerals and the quartz grains were derived from the same source as the extremely fine-grained clay suggests that the parent rock underwent thorough decay before the products of weathering were removed.

The unusual features which characterize the Edwin clay are due to special conditions prevailing at the site of the Jones Butte deposit. A cove in the andesite ridge enabled the concentration here of materials derived from the weathering products of the greenstone. At the same time the quiet waters of the cove did not receive the currents from the east carrying most of the quartz grains and heavy minerals but acted as a settling basin chiefly for the fine-grained clay.

An important stage in the history of the Edwin clay deposit was the leaching of part of the clay by atmospheric waters carrying organic acids. At least three different influxes can be identified. Most important was the introduction of these solutions during the interval between deposition of the red, hematite-stained clay in the lower two-thirds of the present deposit and the upper, stratified Edwin clay. Waters seeping down through the clay dissolved the iron, restored the white color and refractory properties to much of the lower clay, and caused the cylindrical structures and irregular fracture. The solutions also contained CO_2, and much of the iron was precipitated lower down as siderite. Some of the carbonate is also found in a drill hole farther east, but large amounts of iron were probably washed away.

During successive events the deposit was capped first with lignite at the eastern extremity, then with the typical anauxite-bearing Ione sandstone, and finally after a long time interval with 300 feet of resistant volcanic detritus which has preserved Jones Butte and the Edwin clay deposit in their present form.

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EXPLANATION OF PLATE 5

FEATURES OF THE GREENSTONE, LATERITE, AND EDWIN CLAY

Figure

1.—Partly resorbed quartz phenocryst from quartz diorite porphyry showing characteristic chains and zones of bubbles. × 25. Drill hole No. 9, Jones Butte.

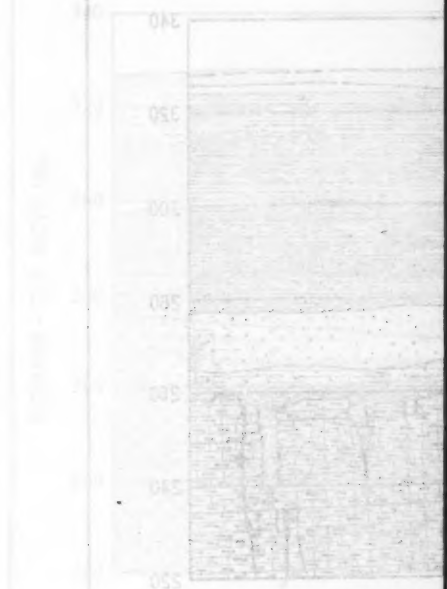
2.—Pisolitic concentrations of gibbsite in ferruginous laterite west of Jones Butte. × 63.

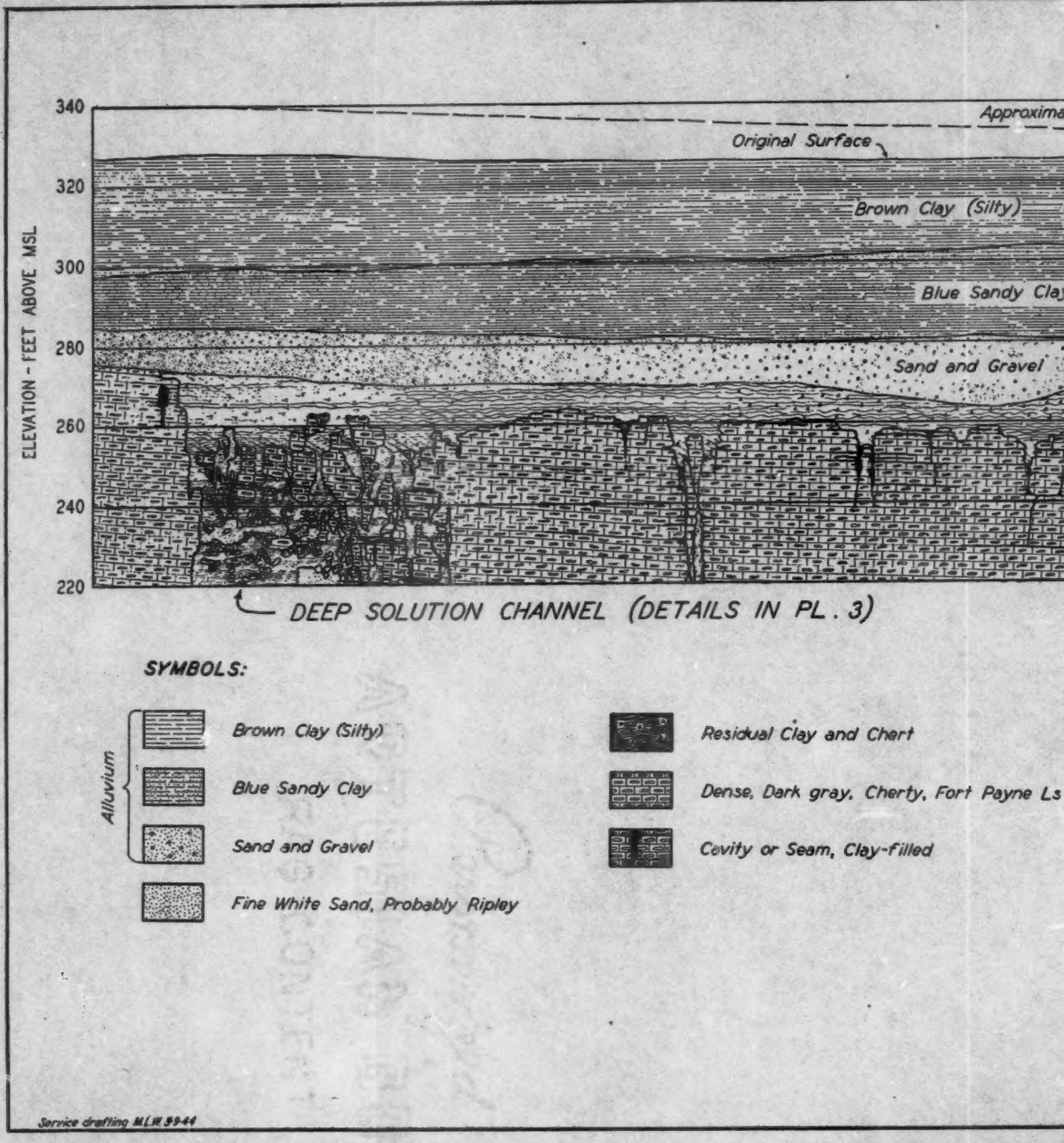
3.—Camera-lucida drawing of amygdule in the andesite. Q, quartz; Ch, chlorite; C, chalcedony; M, matrix. × 136. Woolford pit.

4.—Amygdular structure in the laterite. Q, quartz; Ch, chlorite; G, gibbsite. × 142. Edwin No. 2, Jones Butte.

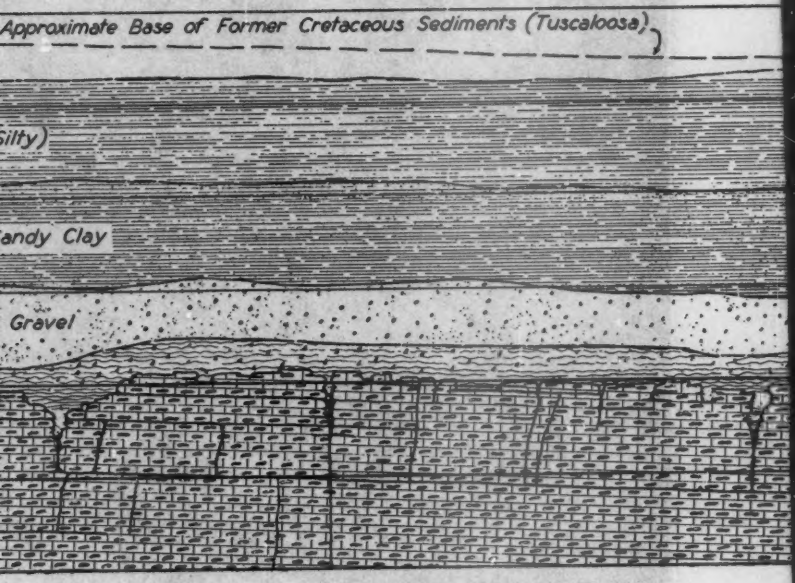
5.—Partly leached Edwin clay with siderite in the hematite-free veins. × 20. Base of lower clay zone, Edwin No. 2.

6.—Siderite replacing quartz grain in hematitic clay shown in figure 5. × 63.





FLOODPLAIN SEDIMENTS OVERLYING THE DISSOLVED BEDROC



LOOKING UPSTREAM

NOTE:
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N 5° W and N 20° W The dip
between 2° SW and 5° SW.

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Normal river level before cons
Dam No. 52 on the Ohio was

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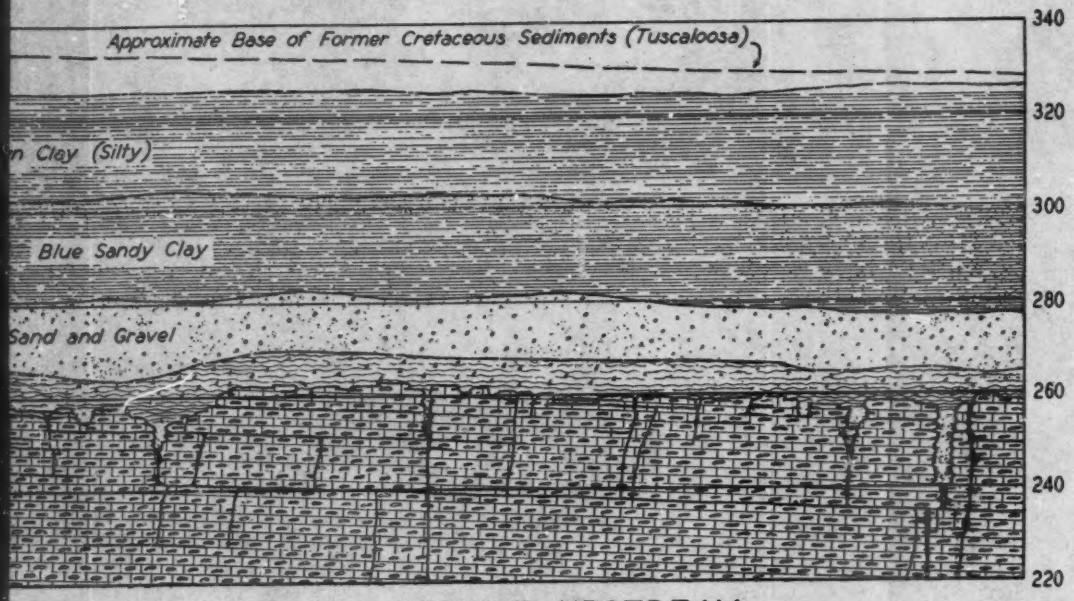
BEDROCK VALLEY FLOOR AT KENTUCKY DAM



LOOKING UPSTREAM

NOTE:
Section not drawn parallel to dip.
The strike of the rock varies between N 5° W and N 20° W The dip varies between 2° SW and 5° SW.

Normal river level before construction of Dam No. 52 on the Ohio was elev. 297±.

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OLVED BEDROCK VALLEY FLOOR AT KENTUCKY DAM

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 56, PP. 39-44 3 PLS., 1 FIG. JANUARY 1945

DEEP SOLUTION CHANNEL IN WESTERN KENTUCKY

BY BERLEN C. MONEYMAKER AND ROGER RHOADES

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ABSTRACT

The limestone and chert floor of the Tennessee Valley at Kentucky Dam is deeply buried by alluvium and extensively dissolved. Solution zones are of two kinds: (1) tabular bedding-plane channels and (2) enlargements of near-vertical joints. Most of this solution has occurred below the water table. Bedrock solution is controlled by (1) lithology (tabular zones are in limestone layers, and vertical zones widen in limestone and narrow in chert layers) and (2) structure (initial ground-water circulation was through joints, and the vertical solution channels are aligned in conformity to joint patterns). Widening of vertical solution zones results from lateral extension along limestone layers and the collapse of intervening chert layers. The jumbled character of the residual chert and clay filling is a reflection of this mechanism. One major vertical solution channel, formed by the coalescence of enlarged vertical joints along the flank of a slightly anticlinal structure, is 70 to 100 feet wide at the bedrock surface (elevation 270) and narrows to extinction at elevation 70. The presence in this solution zone of sand from the Ripley formation, which has since been eroded from the immediate vicinity, indicates that extensive erosion has occurred since solution began—*i.e.*, the solution has been a continuous process of long duration.

INTRODUCTION

Although subriver solution cavities are quite common in the limestones of the Tennessee Valley, most of the known cavities are relatively small (Moneymaker, 1941). During construction of Kentucky Dam and Lock, an unusually large, deep, and extensive solution channel was encountered; it crosses the axis of the dam and

passes beneath the lock immediately downstream from the lower lock gate. Inasmuch as this channel presented foundation problems of unusual magnitude, it was studied in very great detail.

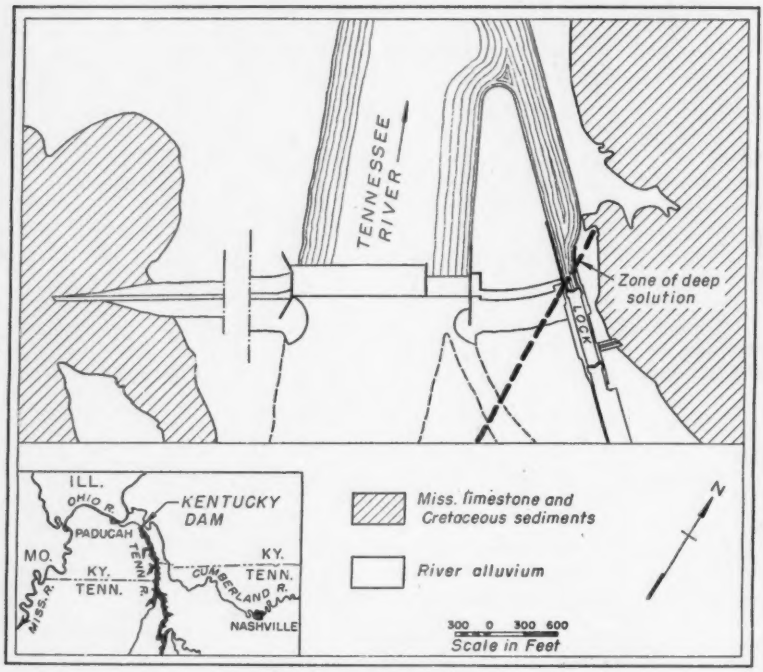

FIGURE 1.—*Location map*

Kentucky Dam[1] is located at Gilbertsville, Kentucky, about 20 miles east of Paducah, on the Tennessee River, 22.4 miles upstream from its confluence with the Ohio (Fig. 1).

ACKNOWLEDGMENTS

The writers gratefully acknowledge their indebtedness to the following members of the Geologic Division, Tennessee Valley Authority: Alvin J. Mistler, Junior Geologist, who, as assistant at the Kentucky Project from July 1937 to June 1941, made most of the observations incorporated in Plate 3; Portland P. Fox, Geologist, who succeeded Mr. Rhoades as project geologist, and served from April 1941 to January 1943; and Leland F. Grant, Associate Geologist, who has been in charge of the geologic work at the Kentucky Project since May 1943.

[1] Kentucky Dam has a total length of 8420 feet and a maximum height of 166 feet. It consists of the following features: concrete gravity spillway, 1176 feet long; navigation lock with chamber, 110 feet wide and 600 feet long; earth embankment west of lock, 740 feet long; earth embankment across the south flood plain, 5681 feet long. The normal head is 57 feet.

GEOLOGY OF THE DAM SITE

All concrete elements of Kentucky Dam are founded on the Fort Payne formation (Mississippian). The lock foundation is about 30 feet below the Fort Payne-Warsaw contact and 515 feet above the Fort Payne-Ridgetop shale contact. The Fort Payne formation, which is strikingly uniform throughout the dam site, consists of thick-bedded, fine-grained, dark-gray to black siliceous limestone containing closely spaced layers, lenses, and nodular masses of black chert.

The geologic structure is quite simple; the strata are nearly horizontal in most places and nowhere are inclined more than 10°. The lock is founded on the southwestern limb of a low, broad anticline with dips of 3°–5° SW. Throughout the dam site the rock is cut by numerous vertical joints, many of which have been enlarged by solution. In the lock foundation northeast-trending joints which strike between N. 15° E. and N. 40° E. predominate numerically, and many of them have been enlarged by solution to considerable depths. Joints trending mainly northwest with strikes between N. 40° W. and N. 50° W. are less numerous and more generally tight. There are many other joints, including those striking about N. 2° W., which have been enlarged enormously by solution.

Very little bedrock is exposed in the lower Tennessee Valley, and this is particularly true at Kentucky Dam. The river flows on a thick deposit of fluviatile sediments; in Pleistocene time it carved a deep valley (not a gorge, as has been stated in some publications) which has been partially refilled with alluvium (Rhoades, 1937). At Kentucky Dam the Pleistocene valley floor is approximately 70 feet lower than that of the present channel.

GENERAL CHARACTER OF BEDROCK SOLUTION

Although the Fort Payne formation consists of impure, siliceous limestone and chert, it is deeply dissolved[2] at Kentucky Dam. The solution has been rigidly controlled by the lithologic character of the rock and the geologic structure, especially vertical joints and bedding planes. The bedding-plane cavities are less numerous but much more extensive than those developed along vertical joints. The largest known cavity developed on a bedding plane ranges in thickness (stratigraphic thickness) from less than a foot to about 6 feet; it is traceable for more than a mile down the dip, and from the top of bedrock at elevation 232 to elevation 56.

The cavities developed along joints are linear channels which extend downward to various elevations. The walls and cross sections of such channels are highly irregular because of the different solubilities of the limestone and chert layers; vertical cavities characteristically are wide in relatively chert free zones and narrow in cherty zones. The same irregularity commonly prevails along their strikes (Pl. 2, fig. 1). Most of these cavities are wholly or partially filled with residual clay and chert. Plate 1 illustrates the corroded nature of the bedrock floor and the general character of the overlying alluvium in the vicinity of the navigation lock.

[2] Solution below the water table is common throughout the Tennessee Valley, as elsewhere. (*Cf.* Moneymaker, 1941.) The water table in the vicinity of Kentucky Dam formerly was depressed to the level of the now-buried bedrock valley floor, which lies at elevation 230 feet, but the solution herein described is mainly below this elevation.

"Bedrock solution in this area follows a genetic pattern which is controlled by the lithology of the limestones: the dissolved zones, starting by the percolation of ground water through vertical joints, originally only minutely open, widen by lateral solution of the limestone beds which lie between the chert layers and lenses. Beginning at elevations near the water table, in the upper parts of the bedrock, this widening takes the form of thin, tabular extensions of the solution cavities outward from the vertical joint, the roof and floor of each tabular cavity being a lens of chert.

"Continuous lateral extension of these tabular openings promotes the collapse of the intervening chert layers, permitting the chert blocks and residual clay of the higher parts of the zone of solution to slump downward and compact, congesting the openings and impeding circulation of ground water through that part of the channel. The parts of the dissolved zones lying above the slumped material become similarly congested by the accumulation of sediments derived from the surface and experience a similar diminution of permeability. Ground water is then able to find an unobstructed channel only at lower levels where the lateral enlargement of the vertical joint has not progressed so far that collapse of the chert roofs and the consequent congestion of the channels has occurred. Continued solution, however, carries this process of lateral enlargement forward until collapse and slumping fills these lower channels, forcing the ground water to seek still lower paths. Thus, progressively, the solution has occurred at deeper and deeper levels. Commonly, slumped and compacted residual chert and clay fill all but the lowest openings, implying that only these lowest openings exist today as channels of unimpeded ground water flow and continued solution. Bedrock solution is therefore apparently occurring today primarily in the lowest parts of the solution channels" (Rhoades, 1941, p. 766–767).

THE DEEP SOLUTION CHANNEL

GENERAL DESCRIPTION

The deep dissolved channel which crosses the axis of the dam near the lower lock gate (Fig. 1) is a linear zone of intense solution trending N. 2° W. at the dam, but curving to N. 7° W. a short distance downstream. It is from 70 to 100 feet wide at the top (elevation 270) and narrows with depth, disappearing at elevation 70, 200 feet below the bedrock surface. Its entire extension along the strike is not known, but drilling indicates that it extends more than 1000 feet and probably more than 2000 feet.

This channel crosses the axis of the dam at an angle of about 50°, and it constituted a serious foundation problem relative to both bearing strength and watertightness. The problem of bearing strength was solved by bridging the dissolved zone with a thick reinforced concrete slab; the problem of leakage was solved by mining out the filling and constructing a concrete cut-off across it at the axis of the main dam. This cut-off is a subterranean arch dam, 195 feet high, built from the top downward (Hays, 1943, p. 1403). In the course of this operation a cross section of the cavity was mapped accurately and studied in great detail (Pl. 3).

Along the line of the cut-off the cavity is 70 feet wide at the top and extends downward without any appreciable narrowing to elevation 217, where it bifurcates, the two prongs tapering to extinction at about elevation 70.

In general, the limiting walls of the channel are highly irregular, the irregularity resulting from selective solution of the limestone-chert series which has developed in the manner described. In some places, however, solution has progressed laterally along a limestone layer to a joint, and the subsequent collapse of intervening chert and limestone layers has left a smooth joint plane as a wall. Continued selective solution eventually restores the irregularity to such walls.

Except for a few relatively small openings near the bottom, the deep channel is entirely filled. The filling consists mainly of residual clay and chert, blocks of limestone, and sand (Pl. 3). The residual chert and clay are derived from the Fort

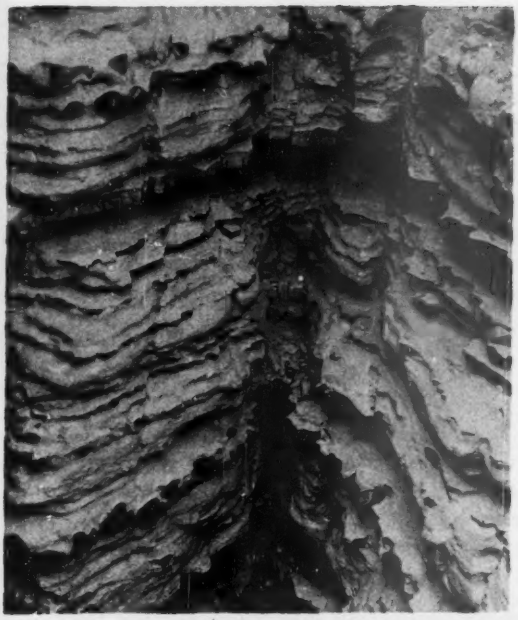

FIGURE 1. TYPICAL IRREGULAR VERTICAL SOLUTION CHANNEL
Narrows almost to extinction along the strike.

FIGURE 2. SLUMPED ROCK AND RESIDUAL CHERT AND CLAY
In zone of solution.

CHARACTERISTIC FEATURES OF SOLUTION CHANNELS

Payne formation and represent the insoluble constituents of the original cherty limestone. This chert and clay constitutes a major part of the material filling both the main deep channel and the associated lateral offshoots. These materials are slumped and intricately contorted due to progressive subsidence as solution continued (Pl. 2, fig. 2). At various elevations (elevations 75 to 125) in the bottom portion of the channel, masses of high-grade limonitic iron ore of delicately stalactitic structure were encountered (Fox, 1942).

Large blocks of undissolved Fort Payne limestone are enclosed in the residual chert and clay. From the top of the channel down to elevations ranging from 230 to 210, the limestone blocks are large and lie nearly horizontally. Many of these blocks are in place; they appear to be attached to and supported by the wall rock. At lower elevations the limestone blocks are slumped and completely enclosed in a matrix of residual chert and clay; they have random orientations and unrelated dips and give the filling a jumbled character.

Pockets and stringers of fine-grained, micaceous quartz sand occur in the upper 100 feet of the channel (down to elevation 160). This sand is almost identical to that of the Ripley formation (upper Cretaceous) in fineness, subrounded character of the grains, color, and mechanical analyses (Rhoades, 1941, p. 769). Remnants of the Ripley formation are present on the adjoining highlands, and undoubtedly the formation once covered the entire area.

ORIGIN AND AGE

The deep channel is in a zone of close jointing developed along the axis of a monoclinal steepening of the dip on the western flank of the anticline upon which the lock is founded. The joints are tensional and tend to converge downward.

The deep channel was never a large open cavity (Moneymaker, 1944, p. 221). It was developed by solution along a series of nearly vertical joints which were widened to form a series of parallel and nearly vertical channels. Each vertical channel developed as described under General Character of Bedrock Solution; lateral enlargement eventually brought about their coalescence and the loss of their individual identities.

Although there is no means of dating the inception of solution, the presence of Ripley sand in the channel filling affords a criterion of the minimum time involved (Rhoades, 1941, p. 769). The deep valley of the lower Tennessee, now partially refilled with alluvium, was cut in the late Pliocene or early Pleistocene; this erosion removed the Ripley formation from the immediate area. Therefore, the sand was introduced into the channel from the surface prior to this time. The lowest unslumped deposit of Ripley sand in the solution channel is at elevation 200, and advanced solution must have progressed to this elevation by early Pleistocene or late Pliocene time. As this elevation is about 245 feet below the former base of the Ripley sand, solution probably was inaugurated in the Miocene or even earlier. Solution may have started very much earlier, possibly during the period of widespread deep weathering which resulted in the formation of the enormous accumulations of residual chert and clay which characterize this section of the country. This period of weathering was in the Mesozoic, prior to the upper Cretaceous period.

Ground-water circulation (hence solution) was still occurring in the small unfilled lower portions of the solution zone until artificially cut off by the concrete bulwark.

CONCLUSIONS

The deep channel at Kentucky Dam throws some light on several aspects of limestone solution: (1) It furnishes additional definite proof that extensive solution of limestone may occur below the water table. Although the water table was depressed to elevation 230 in the Pleistocene, most of the channel has been permanently below the water table. (2) It demonstrates that extreme solution may occur even in impure limestones under conditions involving favorable structure, a continuous supply of moving water, and a long interval of time; chemical purity of the rock becomes relatively unimportant under these conditions. The Fort Payne formation contains nearly as much chert as limestone, and even much of the limestone is impure. (3) The deep channel indicates that long periods of geologic time are represented in the development of large caverns; solution in this area has been a continuous process of long duration.

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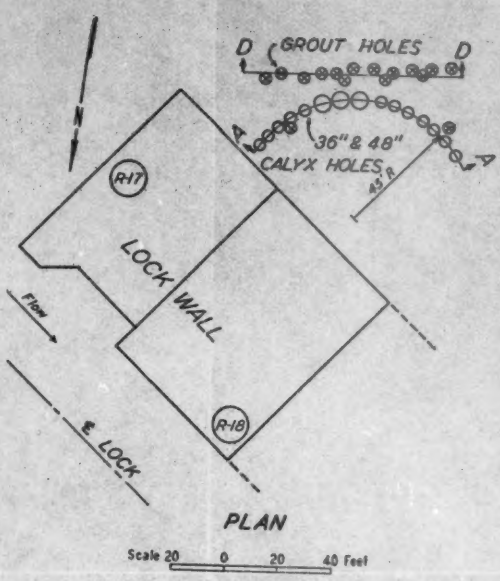

PLAN

Scale 20 0 20 40 Feet

SYMBOLS:

- Cherty limestone - Fort Payne formation
- Residual chert and clay
- Concrete slab
- Grout (neat cement) many thin seams are exaggerated in thickness, but many small areas of grout are not shown.
- Sand (probably from the Ripley formation)
- Diamond drill hole
- Iron ore (Limonite)
- Bedded chert

NOTE:
After removal of flood plain sediments (Pl. 1), a concrete slab was placed on bedrock. By successive steps, working down from the top, the zone was mined out through calyx holes and backfilled with concrete.

Scale 5 0 5 10 15 20 25 Feet

ELEVATION - FEET ABOVE MSL

270 260 250 240 230 220 210 200 190 180 170 160 150 140 130 120 110 100 90 80 70 60

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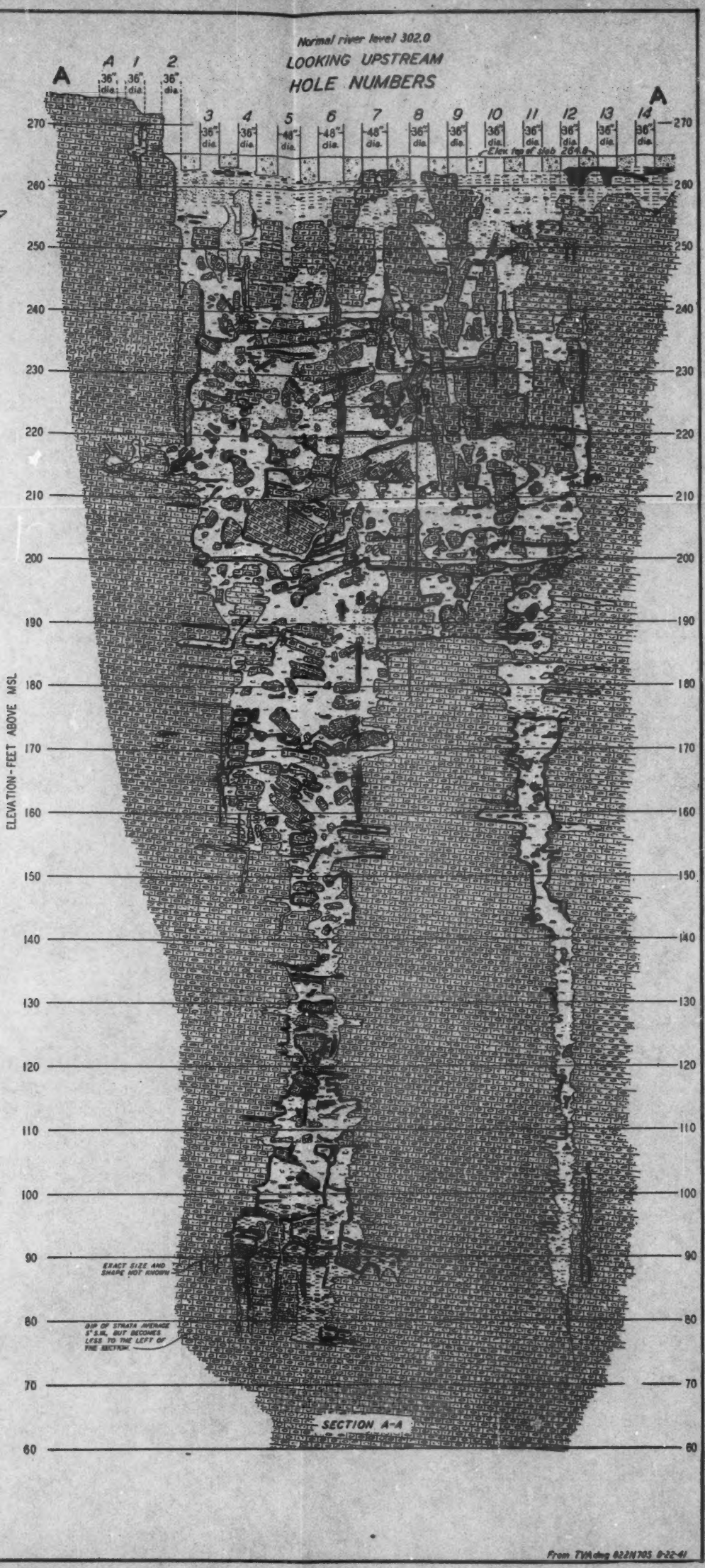

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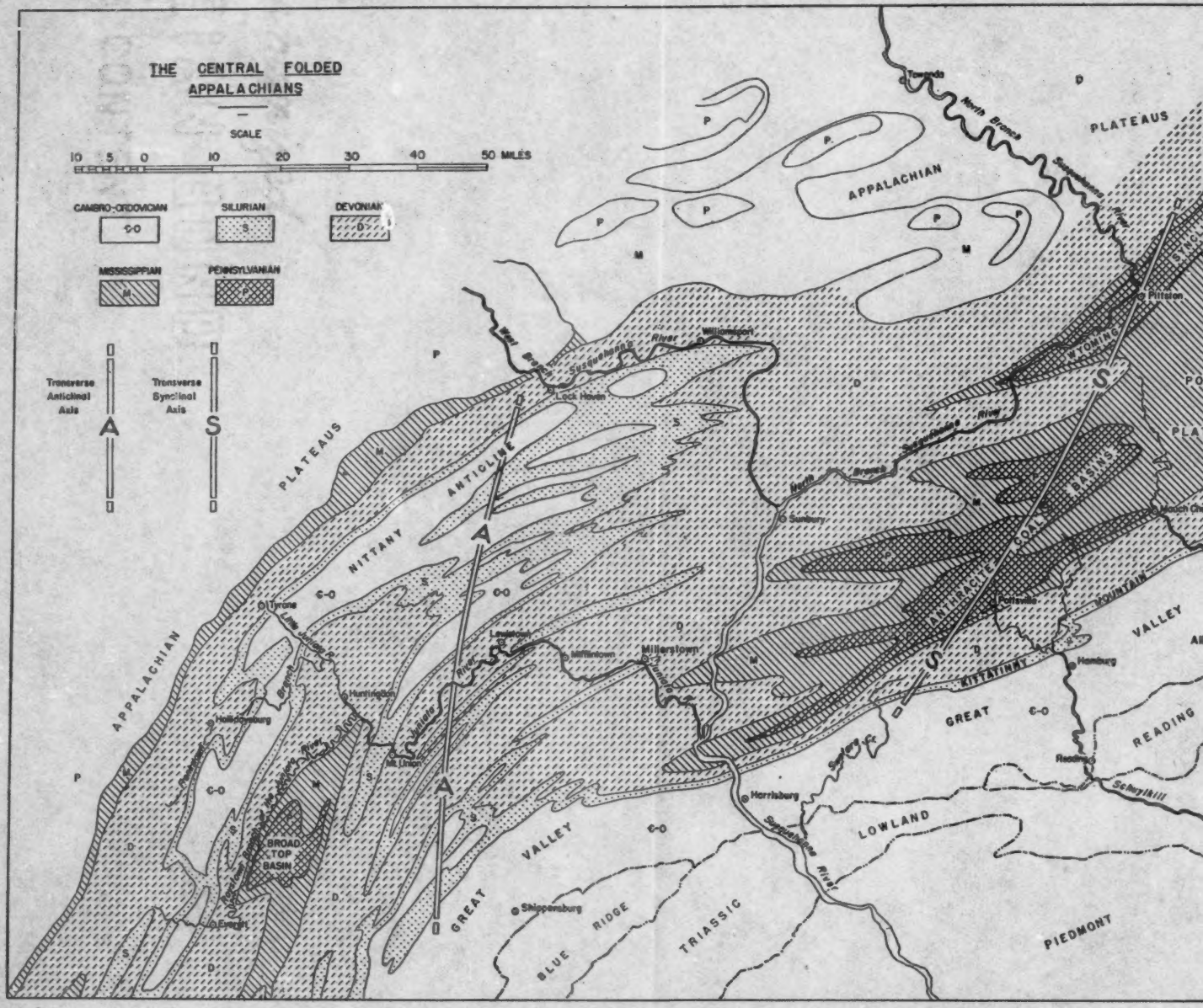
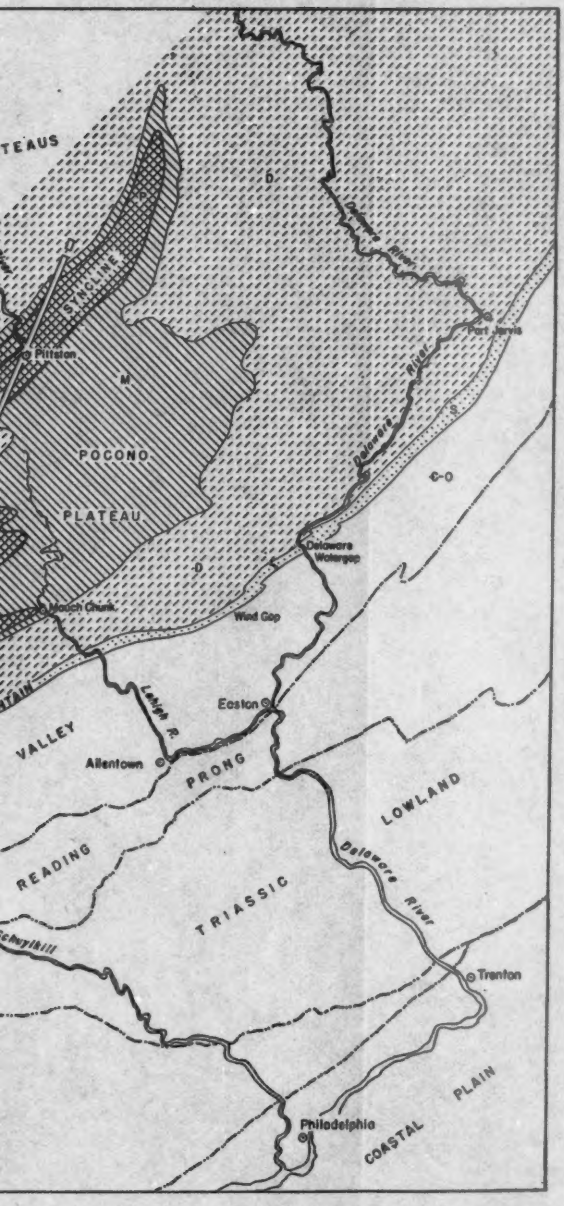


GEOLOGIC MAP OF THE FOLDED APPALACHIANS OF PENNSYLVANIA


TEAUS
SYNCLINE
Pittston
POCONO
PLATEAU
Delaware River
Port Jervis
Delaware Water Gap
Mauch Chunk
Wind Gap
Easton
Lehigh R.
Allentown
VALLEY
PRONG
READING
TRIASSIC
LOWLAND
Delaware River
Trenton
Philadelphia
COASTAL PLAIN

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HYPOTHESES OF STREAM DEVELOPMENT IN THE FOLDED APPALACHIANS OF PENNSYLVANIA

BY ARTHUR N. STRAHLER

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ABSTRACT

New evidence is brought forward in a critical discussion of four hypotheses of origin of southeast-flowing streams in the Pennsylvania folded Appalachians. Development of subsequent streams along transverse faults is concluded to be an unsatisfactory explanation because a study of major water gaps revealed no evidence of faulting and because the few recognized important transverse faults are not the sites of water gaps.

The hypothesis of southeast-flowing Permian consequent streams, modified throughout a long erosion history, is not supported by a study of the positions of present streams with respect to geologic structures. Field evidence fails to sustain certain postulated processes of local superposition. The reversal of a considerable amount of Pennsylvania drainage is not explained.

The hypothesis of progressive piracy and local superposition employs a process of water-gap formation which is not accepted because the numerous wind gaps in the Pennsylvania folded belt show that abandonment of gaps, rather than their production, is the normal process now active. The hypothesis does not account for abundant wind gaps, for repeated alignment of wind and water gaps, or for dual water gaps maintained close together in the same ridge.

Additional evidence supporting the hypothesis of regional superposition is the seeming absence of faults in major water gaps and the failure of streams to utilize the few known fault lines breaking major ridges. Accumulated data invalidate various objections which have been urged against the hypothesis.

INTRODUCTION AND ACKNOWLEDGMENTS

Several explanations of Appalachian drainage have been proposed since Johnson (1931) stated the hypothesis of regional superposition. The present paper brings forward new evidence bearing on the general problem of drainage development in this region and applies it critically to several proposed explanations.

The principal field work has been a study of water gaps in the folded Appalachians of Pennsylvania. The prevalence of transverse faults in major water gaps and, conversely, the extent to which recognizable transverse fault lines have been utilized

as the sites of water gaps have provided data useful in evaluating two of the hypotheses under consideration.

Other evidence presented concerns the relation of Pennsylvania rivers to synclines and transverse downwarps which initial Permian consequent drainage may have followed. To determine more effectively the practicability of a process of stream superposition advocated in one hypothesis data have been compiled on the relation between wind and water gaps of Kittatinny Mountain and the structure of the ridge itself.

Another approach to evaluation of the various hypotheses has been to determine whether certain advocated processes of water-gap development accord with generally accepted principles of stream erosion. Used judiciously and supplemented by available facts, this deductive method is believed to be of value where processes under discussion cannot be observed today. The problem thus treated is whether water gaps were produced by superposition from thrust sheets and overturned folds or by headward growth of streams across barriers of resistant rock.

A number of objections to the hypothesis of regional superposition as outlined by Johnson have been raised in papers in which more recent explanations are proposed. These objections are briefly answered.

This paper treats a part of the central folded Appalachians, or Ridge and Valley Province (Fenneman, 1938, p. 122–123), lying almost entirely in Pennsylvania (Pl. 1). The Potomac River system has commonly been included with the southern Appalachians and is not discussed here. For a region already so much studied it is felt unnecessary to describe the location or physical setting of many detailed features referred to. Most of the area is covered by topographic maps of the U. S. Geological Survey on the scale of 1:62,500.

The writer wishes to acknowledge the aid given by the late Professor Douglas Johnson, not only throughout the course of the investigation, but also in reading and criticising the manuscript.

Acknowledgment is also due Professor A. J. Eardley of the University of Michigan, Professor Frank J. Wright of Denison University, and Professors Walter H. Bucher and Marshall Kay of Columbia University for critically reading the manuscript. To members of the Graduate Seminar in Geomorphology at Columbia University from 1939 to 1941 the writer is grateful for helpful suggestions.

HYPOTHESIS OF TRANSVERSE SUBSEQUENT DRAINAGE

INTRODUCTION AND PREVIOUS WORK

The possibility that southeast-flowing streams of the Pennsylvania folded Appalachians are subsequent along transverse fault lines has been quickly dismissed by some authors but has been made an important part of the regional drainage history by others. The hypothesis deserves detailed study because it seems to provide a simple but effective explanation for the relatively straight, southeast-flowing stream segments which cross the resistant rock ridges through groups of aligned water gaps.

Perhaps the first notable reference to the subject is by Rogers (1858, p. 895–896):

"Two classes of dislocations abound in all zones of plicated and undulated strata.... By far the most numerous, though the shortest and least conspicuous class, are the breaks or faults which run approximately transverse to the strike of the anticlinal and synclinal axes. These may be extensively recognized in the Appalachians, where they are a primary cause of the deep ravines or breaches through the ridges, which furnish passage to nearly all the rivers, and even lesser streams, which drain this chain.... The other far more conspicuous class of dislocations... are the great longitudinal ones."

From this the reader might suppose that the question had long ago been settled, so confident is Rogers' statement. But if we read on we see that he dealt with assumptions rather than with demonstrated facts:

"... parts of the Appalachian chain... embrace many beautiful examples of the deep notches or gaps in the mountain-ridges, above alluded to as indicative of transverse breaks or dislocations in the rocks."

Thus Rogers used the presence of water gaps as evidence of faulting, on the false assumption that there was no other explanation for them. We know today that single water gaps may be produced in a number of ways, without faults being present. Sound independent evidence of faulting necessary to prove Rogers' thesis is quite elusive, as he then notes:

"By far the larger number of these transverse ravines in the great ridges betray no perceptible horizontal displacement or heave of the rocks, though it is possible that a trivial amount of such actual dislocation does exist at nearly every one of them. In some of them, however, the break is very discernible, and in a few instances it becomes a conspicuous feature in the scenery."

Specific cases mentioned by Rogers are the Delaware Watergap, Lorberry Creek gap in Sharp Mountain, West Branch Schuylkill gap in Sharp Mountain, and oblique faults in Canoe, Lock, and Stone mountains. For most water gaps, Rogers' assertion that transverse faults determined the location of stream courses is based upon a false assumption. Treatment of the same subject in the last few years has been scarcely more detailed than that made by Rogers nearly a century ago, and it is surprising that there is no reference to his discussion of transverse faults in any of the papers treated in the following pages.

Peschel (1883, p. 154–156) stated that transverse valleys of the Delaware, Susquehanna, and Potomac rivers in the folded Appalachians are of tectonic origin and existed before occupation by the rivers. Davis (1909, p. 417) criticised Peschel's views:

"Peschel examined our rivers chiefly by means of general maps, with little regard to the structure and complicated history of the region.... There does not seem to be sufficient evidence to support this obsolescent view, for most of the watergaps are located independently of fractures; nor can Peschel's method of river study be trusted as leading to safe conclusions."

The elaborate drainage history then outlined by Davis is based on his belief that transverse fractures have had little, if any, control over the present drainage system.

Hobbs (1904, p. 496) noted the rectilinear courses of many southeast-flowing streams of the northern Appalachian region and suggested that they are subsequent streams developed along a system of fractures. Johnson (1931, p. 33) replied that not only does the lack of evidence of major faults or other fractures coinciding with the stream lines militate against Hobbs' explanation, but, because other explanations

of direct southeast drainage are available, the rectilinear stream courses are not in themselves evidence of lines of structural weakness.

Ashley (1935, p. 1406) again advanced the idea that some water gaps are located upon faults. His discussion is devoted largely to cyclic erosion features of the central Appalachians, and the relationship between water gaps and faults is introduced to support the hypothesis of considerable lowering of ridge crests. A drainage history in which transverse faults have had a controlling influence is nevertheless implied and has been adopted by Meyerhoff and Olmsted (1936, p. 32), who quote from Ashley's paper.

Ashley (1935, p. 1406–1407) states that points of weakness largely due to faulting coincide with Cumberland Gap, Delaware Gap, Lehigh Gap, Schuylkill Gap, and the Susquehanna River gaps in Kittatinny and Second mountains. He intimates that this is evidence against chance superposition of streams from a coastal-plain cover but he does attempt to reconcile Johnson's theory with the coincidence of streams and faults by stating that, if many streams were superposed, those which were let down upon transverse fractures would be favored in the competition that followed. The favored streams, although in some cases small, were able to exploit their advantage, and the final result would be predominant occupation of fracture lines by major streams.

In later papers Ashley (1939; 1940) presents new evidence that the regional superposition theory is untenable. He states as objections that (1) more than a third (72) of 183 water gaps have north-running streams, and (2) that 41 of these gaps are at points of weakness in rocks forming the ridges. Few specific cases of faulting at water gaps are mentioned. In addition to the six cases described in 1935, Ashley (1940, p. 19) refers to several small streams which pass through gaps in Bald Eagle Mountain, the north limb of the Nittany anticline. He notes that the topographic maps show a lack of symmetry between opposite sides of these gaps.

The above evidence is intended to show that there exists between streams and transverse zones of weakness a correlation too strong to have resulted from chance superposition of the streams. Of the 41 gaps, only 11 have been mentioned by name, and for only 6 of these is evidence for faulting presented. In view of this, the statement that about a third of the gaps occupy points of weakness in the ridges is not convincing. Assuming for the time being, however, that 41 out of 183 water gaps do lie at weak points in the ridges, it cannot properly be concluded that this is not the result of chance superposition. In an area of strongly folded rocks regional compression may have produced countless minor fractures, and superposed streams must be expected to encounter a number of these.

Ashley's argument also assumes that under the hypothesis of regional superposition all water gaps must have been cut by streams superposed through the coastal-plain cover. This assumption, treated more fully later, is unwarranted because extensive drainage changes have taken place since superposition occurred. Many gaps cut by smaller streams on limbs of single anticlines or synclines could be products of later resequent or obsequent stream development. Thus only gaps occupied by major southeast-flowing streams can be regarded as critical.

OPPOSED HYPOTHESES AND THEIR DEDUCED CONSEQUENCES

Two major hypotheses will be considered: (1) A marine transgression wiped out Pennsylvania Appalachian drainage. From the uplifted coastal-plain mantle southeast-flowing consequent streams became superposed across underlying structures. After superposition, normal processes of denudation of folded mountains modified the drainage. This hypothesis includes both the regional superposition history outlined by Johnson (1931, p. 14–22) and its later modification by Von Engeln (1942, p. 339–368). (2) Eastern Pennsylvania drainage evolved from initial Permian drainage through a series of changes conditioned by underlying structure and by uplift and warping of the land surface. Thus generalized, this statement may include the hypotheses of Davis (1909), Ashley (1935; 1939; 1940), Meyerhoff and Olmsted (1936), and Thompson (1939). The specific part of this general hypothesis with which we are here concerned will be referred to as the hypothesis of transverse subsequent streams; emphasis is placed on possible occupation of transverse fractures by subsequent streams.

Under the explanation of regional superposition the major southeast-flowing streams, including parts of the Delaware, Lehigh, Schuylkill, Susquehanna, and Juniata rivers, postulated to be superposed consequents, should show only a chance relationship to any faults which break major ridges of the folded belt. A high correlation between water gaps of these major streams and transverse fault lines would be strong evidence against regional superposition only if the total number of fault lines in the region is so small that chance superposition alone could not explain the high proportion of coincidences.

The minor streams which flow southeastward are not stated by Johnson to be superposed. Some of them may have formed as resequents or obsequents upon anticlinal or synclinal structures uncovered after stripping away of the coastal-plain mantle. These might occupy any transverse faults which might be present. Thus the shorter southeast-flowing streams are not of critical importance.

Most of the northwest-flowing streams are small and have much the same significance as the small southeast-flowing ones. Under the explanation of regional superposition the former may have originated as obsequents on northwest-facing scarps of coastal-plain cuestas, later becoming superposed across underlying bands of resistant rock. Again, they may have been subsequent streams flowing in intercuesta lowlands in the coastal-plain beds and have been let down obliquely across resistant strata. Later development of obsequent and resequent streams after removal of the coastal plain may be responsible for many of the northwest-flowing streams. This composite group of streams is therefore not of critical importance.

Two deductions seem to follow from the hypothesis of regional superposition: (1) The major southeast-flowing streams should bear no genetic relationship to transverse faults in the region, but chance coincidence may occur if transverse faults are numerous, and the number of coincidences may be very high if transverse faults are very numerous. (2) The minor streams, flowing either southeast or northwest, are not critical and may, or may not, show a genetic relation to transverse faults.

Under the second general hypothesis—that drainage evolution has been more or less uninterrupted since the mountains were folded in Permian time—most of the

streams should bear a direct relationship either to the initial land surface or to structures in underlying rocks. Subsequent streams would have had a much longer time in which to develop than under regional superposition, thereby increasing the probability that faults would be utilized. Particularly is this so in variations of the general hypothesis which call for drainage reversal from a northwest-flowing Permian consequent or antecedent system to the present southeast-flowing system. The latter group of streams, pushing the divide northwestward across the folds, would tend to exploit whatever fault lines might be present.

Validity of a test of the two explanations rests in large measure on the total number of transverse faults. If they are very few, the headward extension of streams across the ridges as postulated by Thompson (1936; 1939) and Ashley (1940, p. 18) might have had to take place at many points along the ridges where there were no such weaknesses. Under these same structural conditions, chance superposition of streams from a coastal-plain cover would likewise result in streams cutting through ridges at many points where the rock was unbroken.

If, on the other hand, transverse faults were extremely abundant, it would be likely under the hypothesis of continuous development since Permian time that almost all the transverse streams would be located along fault zones. Under the hypothesis of regional superposition the same structural conditions would permit superposed streams to encounter one or more fractures at almost all points where they cut into the ridges.

Only if the faults were moderately abundant—sufficiently numerous to give Permian streams opportunity to gain headway through them to the disadvantage of competitors not favored by fractures, but at the same time sufficiently rare to make coincidence of superposed streams and fractures an infrequent occurrence—would there be any possibility of discriminating the two histories by utilizing this type of evidence. A satisfactory test will require determination of the number of transverse faults in the folded belt.

VALIDITY OF THE ASSUMPTION UNDERLYING THE HYPOTHESIS OF TRANSVERSE SUBSEQUENT STREAMS

Throughout the papers referred to, and so far in the present discussion, the hypothesis of transverse subsequent streams has been allowed to rest on the unquestioned assumption that streams readily grow headward through ridges at places where the resistant strata are weakened by transverse faults.

If a divide is located on a ridge of resistant rock between valley lowlands being eroded on very weak rocks, and if the resistant ridge maker is broken by a transverse fault of moderate displacement (conditions normal for the Pennsylvania folded belt), the weak rocks on opposite sides of the ridge will not be brought into contact between the offset ridge ends. Neither will the resistant stratum at the fault line normally be so badly crushed as to make it as weak as the valley-forming strata.

Reduction of the ridge crest will be slow and will be accomplished by weathering, mass movements of weathered debris, and unconcentrated run-off or minor rill action. No matter how great the river system on one side of the divide, or how

small that on the other side, attack by the destructive forces in the immediate vicinity of the divide, being determined solely by local conditions, will be substantially equal on both sides of a ridge rising above adjacent lowlands. The lowlands will be reduced more rapidly, both because of weakness of the rocks and the concentration of drainage into actively eroding streams. Of two headwater branches of equal length, one flowing transverse to the structure and heading against a divide on resistant rocks, the other flowing parallel with the structure on weak rocks, and heading against a divide in the weak rock area, the latter is far more likely to grow headward rapidly. This is why Appalachian drainage is so predominantly subsequent.

As the divide on the resistant ridge crest is gradually lowered weathering, mass movement, rain wash, and rill action will be more effective upon the crushed zone of a fault than upon adjacent unfractured rock. A sag or col will thus be produced. As the col deepens debris from adjacent walls will increasingly obstruct the depression and progressively retard col development. The narrower the crushed zone, the more quickly will this retarding action become effective. As attack on the weak zone is substantially equal on both sides of the ridge, the ridge-crest divide will continue to cross the col not far from its center. If the crushed zone is very wide and very weak, the col may become relatively broad and deep. But the drainage divide will still lie in the col so long as the weak-rock belts on either side of the resistant ridge are being lowered more actively than the col itself. Even when lowering of the weak-rock belts has become relatively slow or has ceased, the divide can be transferred from the col to either lowland, with resulting capture of lowland drainage, only under conditions that are rare or nonexistent in the Pennsylvania Appalachians. These conditions are discussed later.

Thus, even though a fault zone in a resistant ridge constitutes a weakness sufficient to produce a col, it does not necessarily follow that a tributary of the more active drainage system can readily push through the ridge at such a point to ramify throughout the lowland beyond. The continuity of subsequent lowlands and the abundance of wind gaps in resistant ridges provide concrete field evidence that longitudinal streams have grown at the expense of transverse streams, not the reverse.

Special conditions might produce water gaps by headward growth of streams through a resistant ridge. If the ridge is so long that it is virtually impossible for subsequent streams to work around it to capture the lowland beyond, and there develops a very great difference in elevation of weak-rock lowlands on both sides of the ridge, the divide may in time become highly asymmetrical and resemble an escarpment. That slope of the divide facing the lower lowland, being much steeper than the other, would retreat more rapidly, because of the faster removal of weathered material by mass movements and unconcentrated runoff. The divide could then be pushed across the resistant strata to the weak-rock area beyond. This process would operate regardless of whether any faults break the ridge, although such structures would facilitate and localize it. Because of the requirement of extremely long ridges, this method of water-gap formation could have but limited application in the Pennsylvania folded belt. Kittatinny Mountain alone seems to possess the

requisite length, and there is no evidence that the lowland north of that ridge ever had an elevation greatly in excess of that to the south. On the contrary, the numerous wind gaps indicate that transverse streams already established across the ridges lost drainage to longitudinal subsequent streams.

Where the resistant stratum is so greatly offset that weak-rock belts on opposite sides of the ridge are brought into contact along the fault line, a subsequent stream might gain access to the lowland beyond. This situation really involves two distinct ridges separated by a weak-rock lowland and can have no bearing on the cases actually under consideration. In these, the offsets, whether real or assumed, are relatively slight and do not destroy the continuity of the ridge.

One could, perhaps, imagine a fault zone so broad, in which the resistant formation was so thoroughly crushed, as to render it fully as susceptible to removal as the weak shales and limestones of subsequent lowlands leading around the ends of pitching anticlinal and synclinal mountains. No effective barrier would exist, and a transverse stream might gain access to the subsequent lowland beyond. No evidence of any transverse fault zone of this type has been found in the Pennsylvania ridges. Rapids, often present in Appalachian water gaps, indicate that the rock is there more resistant than in the lowlands upstream and downstream from the gap.

It is concluded that the concept of water gaps forming by headward growth of streams across ridges along transverse fault zones is of highly doubtful validity except under the most unusual circumstances, and that any hypothesis of Appalachian drainage based on this concept must be viewed with great caution. The question as to whether transverse faults are present in the folded belt of Pennsylvania, and whether any of them are related to major transverse streams, must next be considered. Lack of genetic relationship between observed transverse faults and water gaps will tend to support the conclusion stated above.

The possibility that tensional fissures or major joints produced during the folding have had an effect on drainage should be considered. If the anticlines were broken by large transverse joints, the drainage might be localized. Because such fractures would be produced without offsetting the strata, they would be difficult or impossible to detect after occupation by a stream. In view of the difficulties already discussed, of water-gap development along fault lines where the rock is actually crushed, a joint seems too small a structure to cause a water gap. Furthermore, it does not seem likely that a single joint would extend across two or more anticlines to produce the aligned groups of gaps.

TOPOGRAPHIC EVIDENCES OF FAULTING

Inasmuch as offsetting of Appalachian ridge crests at water gaps has been accepted by some as evidence of transverse faulting, it is desirable to recall that other conditions produce such offsetting.

Lowering of the ridge crest on one side of the gap more than on the other will cause the crests to be offset at the gap, provided the beds are not vertical. The lower the angle of dip, the greater will be the offset per unit difference in elevation of the two ridge crests. Lowering of the ridge crest on one side of a gap more than on

the other may be due to (1) lateral shifting of the stream as it cut the gap, or (2) undercutting of the flank of the ridge by the main river or by a tributary (Fig. 1).

Minor warps and flexures in ridge-forming strata may give rise to offsetting of the ridge crest. After a segment of ridge has been removed to form the gap a

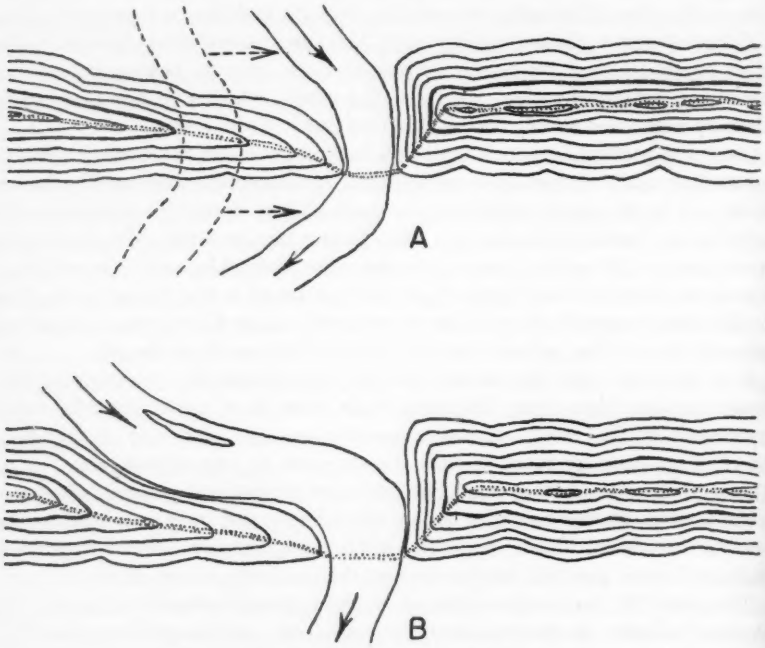

FIGURE 1.—*Offset forms of ridge crests at water gaps*

Brought about by shifting of the crest down-dip on one side of the water gap. In (A) the western ridge, representing the slipoff slope, has been lowered by the river shifting laterally toward the outside of a bend. In (B) the lowering has resulted from undercutting by the river on the north side of the ridge. Outcrop of resistant stratum shown by stippled line.

straight-line projection of ridge crests to the gap center will reveal an offset which might erroneously be attributed to faulting (Fig. 2).

In view of these possibilities it is dangerous to assume faulting in a water gap merely on the basis of offset ridge crests. Tracing strata across the gap seems to be the only positive means of determining whether offset ridge crests are a result of faulting, unless the offset is so great that simple warping or differential lowering of the ridge crests will not account for it.

WATER GAPS OF THE MAJOR STREAMS

Introductory statement.—As pointed out, the major southeast-flowing streams alone are of critical importance in testing the hypotheses. These streams include parts of the Delaware, Lehigh, Schuylkill, Susquehanna, and Juniata rivers. Most

of them cross two or more major ridges through a series of roughly aligned water gaps. The 24 gaps cut by these streams through ridges of Tuscarora, Pocono, and Pottsville strata were examined by the writer (Pl. 2).

Susquehanna River gap in Kittatinny Mountain (Harrisburg, Pa., quadrangle).—Kittatinny Mountain (Blue Mountain) in the vicinity of the Susquehanna River is

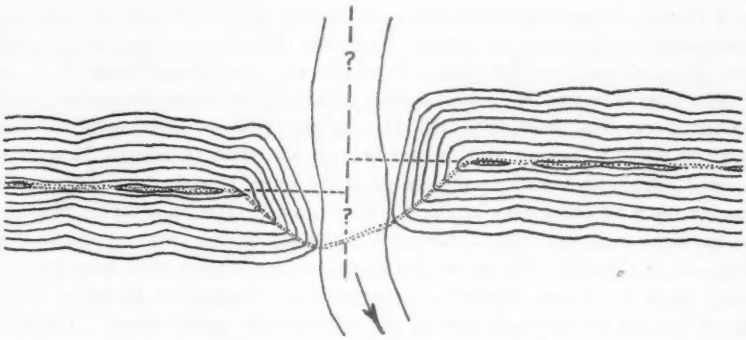

FIGURE 2.—*Offsetting of ridge at a water gap*

Caused by flexing only. An erroneous interpretation of faulting is indicated, while the true outcrop of the resistant stratum is shown by a stippled line.

upheld by the Silurian Tuscarora and Rose Hill sandstones. The Tuscarora formation is here relatively thin (275 feet), and the ridge crest itself is upheld by the Rose Hill beds (750 feet). The formations are overturned and dip about 70°S. The highly resistant Tuscarora sandstone may be traced across the river channel as a continuous and apparently unbroken bed, and thence up both walls of the water gap. Strike readings on the Tuscarora bed on both sides of the gap and in the river bed agreed with each other and with the general strike of the ridge as determined from the topographic map. Thus the evidence seems conclusive that strata of the ridge are essentially unbroken in the vicinity of the gap and that no significant transverse fault zone is present.

Rogers (1858, p. 896) reported a transverse dislocation in this gap but gave no evidence. Ashley (1935, p. 1406) states that faulting is not seen, but that a detached block of Oriskany sandstone on the west bank, with no counterpart on the east bank, and much minor faulting suggest a fault zone. He seems to be referring to a paper by Willard (1931) in which the position of a block of Oriskany sandstone atop Little Mountain, just north of Kittatinny Mountain, is attributed to thrust faulting. Cloos and Broedel (1943) believe the supposed Oriskany is Montebello sandstone and that the Oriskany is absent due to lack of deposition. Moreover they disagree with Willard's interpretation of the nature and magnitude of the supposed thrust faults.

The minor faulting referred to by Ashley seems to be that described by Cloos and Broedel (1943) in the road cut on the west bank of the river. Here the Tuscarora is broken by numerous gently dipping reverse faults cutting the bedding at right angles and having displacements from a few inches to 3 feet. The faults seem to

strike about parallel with the strata and are therefore not so oriented as to produce transverse weaknesses in the resistant beds, nor do they suggest the presence of larger transverse faults. It should be recalled that these various minor structural features can be observed only in the walls of the water gap and in all probability are generally present elsewhere throughout the ridge where they cannot be seen. More important than such minor features is the fact that the resistant strata at this gap can be traced continuously across the river without any visible zone of weakness.

Susquehanna River gap in Second Mountain (Harrisburg, Pa., quadrangle).—About 2½ miles north of Kittatinny Mountain gap, the Susquehanna River flows through a gap in Second Mountain, a ridge upheld by the Pocono formation which is overturned and dips about 65°S. (Pl. 2). Ashley (1935, p. 1406) suggested that a transverse fault is located at this gap because the crest of Second Mountain, if projected from both sides into the center of the water gap, shows a distinct offset. However, such an offset may be due to other causes. The topographic map shows that the ridge crest for about a mile west of the gap is lower than at corresponding points east of the gap. The strata dip south, hence the ridge crest must have been shifted south as it was lowered. Corresponding topographic elements on both sides of the gap at the south base of the ridge do not appear offset. The greater lowering of the ridge crest on the west side of the gap is due to undercutting of the ridge along its northern side by the Susquehanna River. This undercutting caused more rapid denudation of the north slopes of the ridge, which in turn caused a greater lowering of the ridge crest.

To demonstrate that faulting is absent from the gap it would be necessary to trace individual beds across the gap, as was done in the Kittatinny Mountain gap. In the Pocono formation no single stratum is continuously exposed across the entire river bed, although some beds can be followed for considerable distances by the eye. Strike readings wherever taken were in close agreement, indicating that the southward deflection of the ridge crest on the west side of the gap is not due to bending of the beds. Thus there is no evidence of any weakening of the Pocono ridge at the gap by either flexing or faulting. If a fault of appreciable displacement breaks the Pocono strata at this water gap it seems likely, though not necessary, that some break would also be found in Kittatinny Mountain and Peters Mountain. It has been shown that none exists in the Kittatinny Mountain gap which is directly in line and only 2 miles away, while the crest of Peters Mountain, 3½ miles away, shows no deflection at all.

Susquehanna River gap in Peters Mountain (New Bloomfield, Pa., quadrangle).—A few miles upstream from Second Mountain gap the Susquehanna River flows through a water gap in Peters Mountain, a major ridge upheld by the Pocono formation and representing the north limb of the overturned syncline of which Second Mountain is the south limb (Pl. 2). Ashley (1935) makes no mention of a transverse fault in this gap. The ridges on both sides of the water gap show distinct if slight differences in form. Ver Steeg (1930, p. 188) noted that a profile of the gap reveals a steeper concave west wall and a more gentle convex slope on the east side. This difference he interpreted as the result of undercutting of the west bank by the Susquehanna River. There are, however, no topographic irregularities which

suggest any break of the strata at the gap. Examination gave no conclusive proof as to whether or not a fault exists because strata could not be traced across the river bed.

Susquehanna River gap in Berry Mountain (Millersburg, Pa., quadrangle).—North of Peters Mountain the Susquehanna River flows across another syncline represented by Berry and Mahantango mountains (Pl. 2). The Pocono strata in Berry Mountain gap dip 65°–75°N. There is no offset of the ridge to suggest a fault. Limited exposures are available in the walls of the gap, and riffles are seen in the river bed, but it was not possible to trace a single stratum across the gap. The conclusion therefore remains in doubt, but certainly no significant fault is indicated.

Susquehanna River Gap in Mahantango Mountain (Millersburg, Pa., quadrangle).—The water gap in Mahantango Mountain is wider than those previously described. The ridge for about 2 miles east of the gap is considerably narrower than elsewhere because the north base of the ridge is undercut by the Susquehanna River (Pl. 2). On the west side of the gap the ridge is split by a small subsequent stream which has apparently developed along a weaker bed in the Pocono formation. Nothing in the topography suggests a fault. Riffles are present in the river bed, and outcrops are available in the walls of the gap, but no individual bed could be traced across the gap.

Summary of Susquehanna River gaps.—The five gaps discussed above form a critical series. Not only is the Susquehanna River the largest stream in the region, but it flows across four ridges of resistant Pocono sandstone although there is, a short distance to the west, a path around these barriers entirely on weak rocks (Pl. 2). It is difficult to imagine what other explanations of this river's course could be devised than the two being treated here—that it was superposed, or that it grew along a fault line where the resistant strata were crushed and rendered as weak as the near-by shales. Only in the Kittatinny Mountain gap can it be proved that the resistant strata are unbroken. For the others there is no suggestion of weakness at the gaps, and the strata appear to pass uninterrupted across the gap. The riffles in the river bed attest that the rock in the gaps is much more durable than the adjacent shales and limestones. Even if it were demonstrated that minor faulting is present, it would be necessary to show that the fault lines constitute zones even less resistant to weathering and stream erosion than the rock which underlies the broad subsequent lowlands to the west. Furthermore, it is not enough to find a suggestion or even proof of faulting in only one or two of these water gaps, since the others would still require a different explanation. Under the hypothesis of regional superposition one or two gaps might by chance coincide with faults. It is therefore concluded that the hypothesis of subsequent stream development along fault lines is inadequate for the Susquehanna River gaps above Harrisburg, but that the hypothesis of regional superposition satisfactorily accounts for the observed facts.

Port Clinton gap of the Schuylkill through Kittatinny Mountain (Pottsville, Pa., quadrangle).—A few miles north of Hamburg the Schuylkill flows southward through Port Clinton gap in Kittatinny Mountain (Pl. 2). Ashley (1935, p. 1406) wrote: "Schuylkill Gap in the same mountain occurs where a fault cuts the mountain obliquely, and the river correspondingly makes an S-turn in crossing the ridge."

Throughout this area Kittatinny Mountain is relatively complex in structure. The Tuscarora and Clinton formations, here the resistant strata, dip about 80°N. at the gap. The nearly horizontal shales of the Martinsburg (Ordovician) formation abut the almost vertical Tuscarora quartzite. Ashley has explained no further than the statement quoted above, but this contact is evidently the fault to which he refers. The Schuylkill does turn and flow along the contact for a short distance, but it is not generally agreed that this is a fault contact. Several authors regard this as an angular unconformity representing the Taconian orogeny which closed the Ordovician (Stose, 1930; Miller, 1926, p. 503–504, 509–511). The same relationship exists farther northeast where deformational structures within the Hudson River shales but absent in the overlying Silurian Manlius strata provide independent evidence of this orogeny (Schuchert and Longwell, 1932). Thus the interpretation as an angular unconformity at Port Clinton gap, rather than as a fault, seems preferable.

Assuming that there is a fault, which represents a line of weakness, there is still good reason to conclude that it does not cut across the resistant formations of the mountain. Field examination supports studies showing that this line of contact, exposed on both sides of the gap, runs parallel with the mountain rather than across it. This relationship is expectable if the feature is an angular unconformity. Nothing in the topography of the Port Clinton gap area suggests that a fault breaks the resistant strata. The gap is asymmetrical because of the curving course of the Schuylkill, but the ridge crests on both sides of the gap fall into line.

Lehigh Gap (Mauch Chunk, Pa., quadrangle).—Lehigh Gap in Kittatinny Mountain is occupied by the southeast-flowing Lehigh River (Pl. 2). This deep, narrow water gap has been fully described by Chance (1882 b) and more recently by Ver Steeg (1930, p. 159–165). Regarding a possible zone of weakness here Ashley (1935, p. 1406) states: "Lehigh Gap cuts through Kittatinny Mountain, where a local bulge in the structures results in differences of structure either side of the river." At the Lehigh Gap the resistant Tuscarora (Medina and Oneida) strata dip about 35°N., but about midway through the gap a small anticlinal flexure, with its axis parallel to the ridge, has resulted in a southward bulging or offsetting of the ridge crest near the gap. Ashley seems to refer to this feature. Its presence, however, in no way indicates a weakness in the resistant strata of the ridge. Similar irregularities characterize adjacent parts of Kittatinny Mountain where no gaps exist, and probably many minor flexures affect the resistant strata. Behre (1933, pl. 25) has mapped Lehigh Gap and the area south in detail but has shown no faulting in or near the gap.

Delaware Watergap (Delaware Watergap, Pa., quadrangle).—This water gap in Kittatinny Mountain (Pl. 2) has been described fully by Chance (1882a) and Ver Steeg (1930, p. 152–159). The ridge is upheld by the Oneida and Clinton (Shawangunk) formations which dip 40°–45°NW. The ridge-crest line, projected from each side to the center of the gap, shows an offset of about 700 feet which is probably the feature Ashley (1935, p. 1406) had in mind when he stated that "the Delaware Water Gap follows a transverse fault with horizontal displacement of 700 feet."

Rogers (1858, p. 896) first noted this offset of the ridge crests. He concluded that a transverse dislocation had thrown the New Jersey side relatively northward several hundred feet. Because other causes might produce such an offset, it is necessary to seek independent evidence of the fault. The resistant strata on either side of the gap are closest together at river level. If a fault is present, corresponding points in the sedimentary series on both sides of the river should not lie along a line

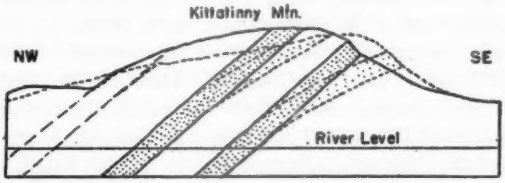

FIGURE 3.—*Structure sections of Kittatinny Mountain at the Delaware Watergap*
In solid lines is shown the section on the northeast side of the gap; dashed lines show the southwest side. The strata are not offset at river level. After Chance (1882).

parallel with the general trend of Kittatinny Mountain, but on a line striking more strongly north of east. Strike readings taken near river level showed no such perceptible offset. A fault of appreciable magnitude in the Delaware Watergap would probably affect Blockhead Mountain, a mile north of Kittatinny Mountain. No suggestion of faulting was found there.

On the northeast side of the gap the strata dip about 40° from river level to the summit of the ridge. On the southwest side, however, although the dip at river level is also 40°, it decreases toward the ridge summit (Fig. 3). Lessening of dip would carry the ridge crest farther southward on the southwest side and thus explain the apparent offset. Slight warping is all that is required to restore the strata across the gap. Warping of this type and of even greater magnitude affects Kittatinny Mountain between the Delaware Watergap and Windgap, where it has produced the Big and Little Offsets (Pl. 2). Thus the offset accepted by Ashley as evidence of faulting is readily explained otherwise.

Schuylkill and West Branch Schuylkill water gaps through Second and Sharp mountains (Pottsville, Pa., quadrangle).—Near Pottsville both the Schuylkill and its West Branch flow across Second and Sharp mountains, ridges of the Pocono and Pottsville formations (Pl. 2). The two streams join about 2 miles south of Second Mountain in which the gaps are only about a mile apart, while those in Sharp Mountain are 2½ miles apart. Johnson (1931, p. 64) considered these gaps to have been formed by superposed streams. It is therefore important to determine whether faults coincide with the water gaps.

Under the regional-superposition hypothesis both streams flowed on the coastal plain in much the same position as today, and after becoming superposed upon the ridges neither was able to capture the other. It seems reasonable to find a few cases of dual superposition above a common junction. Under the subsequent-stream hypothesis it is also necessary to suppose that the two streams maintained such equality of advantages that one could not capture the other. To preserve these conditions transverse fault lines must have been available for both streams; other-

wise one would have had a considerable advantage. Therefore one requisite of the hypothesis of headward growth of a subsequent stream is that faults occur in the gaps of both streams.

The topographic map shows that neither Second Mountain nor Sharp Mountain is offset at the gaps; structural continuity appears to be unbroken. Field examination confirmed this. On the Pennsylvania Geological Survey mine sheets of this region (Ashburner, 1885) corresponding minor topographic features are represented as matching on both sides of the water gaps in Sharp Mountain, while lines showing outcrops of specific stratigraphic horizons pass uninterrupted across the gaps.

Little Schuylkill River gaps at Tamaqua (Hazleton, Pa., quadrangle).—Near Tamaqua the Little Schuylkill River flows southward across three ridges (Pl. 2). The northernmost two, Pisgah and Nesquehoning mountains, are composed of the Pottsville formation. The southernmost, Mauch Chunk Ridge, is upheld by the Pocono formation. The three gaps are in an almost straight line. Under the hypothesis of regional superposition this alignment is explained by superposition of a nearly straight stretch of the former consequent Little Schuylkill River. Under the hypothesis of subsequent origin of the major rivers, the alignment would represent subsequent development along a single fault cutting the three ridges.

The topographic map shows no pronounced offset of the ridges. The gap in Pisgah Mountain shows a slight offset of the ridge crests on both sides. Too much reliance should not be placed on the map, because it was made at an early date and looks like a reconnaissance survey. Field examination did, however, show that the strata are continuous across the gap in Pisgah Mountain. A resistant bed stands out boldly on both sides of the gap with only a narrow break where the river passes through. The strike of a line connecting these outcrops coincided with the general strike of Pisgah Mountain. In the gaps through Nesequehoning and Mauch Chunk mountains no suitable exposures were found because the river curves considerably and talus obscures the slip-off slopes. The evidence in Pisgah Mountain Gap is, however, sufficient to invalidate the subsequent-stream hypothesis here.

Lehigh River gap in Mauch Chunk Ridge-Bear Mountain (Mauch Chunk, Pa., quadrangle).—Near Mauch Chunk the Lehigh River flows southward across Mauch Chunk Ridge-Bear Mountain, a ridge of Pocono sandstone dipping steeply northward (Pl. 2). In passing through the ridge the river follows an S-curve resulting in a long, winding water gap. No evidence of faulting was found in this gap, and there appear to be no topographic irregularities which demand such explanation. Because the river course is serpentine, a correspondingly curved fault line would seem to be called for, a requirement which renders highly improbable the subsequent-stream hypothesis.

North Branch Susquehanna River gap north of Pittston (Pittston, Pa., quadrangle).—The Susquehanna River flows southeastward into the Wyoming Coal Basin through a water gap in a double ridge of Pocono and Pottsville strata which forms the north limb of the syncline (Pl. 2). Irregularity of the topography near the gap and the curving course of the river obscure any topographic evidence of a possible transverse fault. Darton (1940, Pl. 10) prepared a structure map of the Wyoming syncline in which the lowest coal bed above the Pottsville conglomerate is contoured in con-

siderable detail. At this gap, as at all others in the region mapped, no transverse faulting, or even appreciable transverse flexing of the stratum, is shown.

North Branch Susquehanna River gap at Nanticoke (Shickshinny, Pa., quadrangle).—Near Nanticoke the Susquehanna River turns sharply northward through the inner rim of the basin (Pl. 2). After passing through the gap, it turns southwestward in a subsequent valley between the Pottsville and Pocono ridges. From Pittston, where the Susquehanna enters the Wyoming syncline, to Sunbury, where it turns southward, it is for the most part parallel to Appalachian structure and therefore probably subsequent. It is not critical in distinguishing between hypotheses of Appalachian drainage. Deflection of the river at Nanticoke may be due to "structural" superposition during denudation of the region, or it may represent an obsequent portion of an eastward-growing subsequent stream which developed into the present Susquehanna. Also to be considered is Itter's statement (1938, p. 43) that this portion of the Susquehanna is not a simple subsequent stream, but a superposed subsequent initially developed in an inner lowland or inter-cuesta lowland of the mantling coastal plain. The case would resemble that of the former Hudson River crossing and recrossing the Watchung trap ridges (Johnson, 1931, p. 76–131). If there is a transverse fault at the Nanticoke gap, a simple subsequent origin for the North Branch Susquehanna between Pittston and Sunbury may seem to be favored. If no evidence of faulting be found, either explanation is possible, and the case is not critical.

The Pottsville ridge crest is offset at Nanticoke gap, which might suggest a transverse fault. On the other hand, Darton mapped no break across the line of the gap. Furthermore, no corresponding offset occurs in the Pocono sandstone ridge, directly in line northwest of the gap and only a mile away. If the offset ridge at Nanticoke gap represents a significant displacement the near-by ridge should also be affected.

Offset of the Pottsville ridge at Nanticoke gap is easily explained without recourse to faulting. The base of the ridge west of the gap has been undercut by the river and the ridge narrowed. This must have resulted in a down-dip shift of its crest southward with respect to the ridge crest on the east side of the gap.

North Branch Susquehanna River gaps at Shickshinny (Shickshinny, Pa., quadrangle).—At Shickshinny the Susquehanna River turns out of the Wyoming syncline and crosses the end of the synclinal mountain formed of Pottsville strata, then flows across Penobscott Mountain, the Pocono ridge (Pl. 2). In his diagrams of the evolution of Appalachian drainage Johnson (1931, p. 64) represented the North Branch Susquehanna between Pittston and Sunbury as a later subsequent stream. The point to be determined is whether the gaps are located along a fault, in which case the stream may have grown headward as a subsequent to drain a closed synclinal basin at its most favorable point; or whether there is no line of weakness, in which case the stream may here be a normal obsequent gaining entrance into the syncline. There is the further possibility that, as at Nanticoke, the Susquehanna may here be a superposed subsequent, genetically unrelated to faulting even if fractures are present. A fourth possibility is that these gaps, together with the gap of Shickshinny Creek directly in line to the north, were carved by a southeast-flowing consequent stream superposed across the end of the Wyoming syncline and later diverted

above and below that point. The gaps would then be unrelated to any transverse fault zones. No evidence of faulting was found in the Shickshinny gaps, nor are any dislocations shown on Darton's structure-contour map.

Little Juniata River gap at Tyrone (Tyrone, Pa., quadrangle).—At Tyrone the Little Juniata River turns southeastward through a water gap in Bald Eagle Mountain, a ridge formed by the resistant Tuscarora and Oswego sandstones (Pl. 2). Bald Eagle Mountain varies greatly in topographic form on opposite sides of the gap. The ridge west of the gap is broad and double-crested because it has been split longitudinally by a subsequent stream excavating the weaker Juniata red beds from between the Tuscarora and Oswego formations. No evidence of a transverse fault was found at the gap, nor has any been indicated on the geologic map of the Tyrone quadrangle (Butts *et al.*, 1939).

Juniata River and Frankstown Branch gaps in Tussey Mountain (Tyrone, Pa., quadrangle).—From its water gap near Tyrone the Little Juniata River flows southeastward across Nittany Valley which the river leaves through a water gap in Tussey Mountain (Pl. 2). Two miles southwest of this gap Tussey Mountain is cut by another water gap through which flows the Frankstown Branch of the Juniata River. The two streams join a few miles farther downstream. Davis (1909, p. 457–458) thought it unreasonable that two independent gaps could be maintained so close together. Johnson (1931, p. 71–72) stated that if the two streams were superposed from a coastal plain they would normally have maintained their independent gaps, since great inequality of size between the two streams would have been required to permit piracy. The hypothesis that the gaps are controlled by transverse fault lines requires fractures in both water gaps, for if only one lay along a fault zone, the stream thus favored would have quickly conquered the subsequent lowland beyond the ridge, making development of the second gap impossible.

Tussey Mountain is upheld by the Tuscarora formation, dipping 25°–30°SE. The relatively low dip has resulted in the development of "flared" V gaps, constricted at the south end where the resistant beds descend to river level. The formation is well exposed, and no displacement was observed. No transverse faulting was mapped by Butts *et al.* (1939).

Juniata River gap in Jacks Mountain (Mt. Union, Pa., quadrangle).—Near Mt. Union the Juniata River flows southeastward across Jacks Mountain, a great anticlinal ridge upheld by the Tuscarora formation (Pl. 2). Near the gap the anticline is but slightly breached and rises to elevations of above 2200 feet. Under the hypothesis of regional superposition this gap is the third of the series which lies upon a southeast-trending line representing the course of an initial consequent stream on the Cretaceous coastal plain. In addition to the second hypothesis, that the stream is subsequent on a transverse fault zone, there is the possibility that it has been locally superposed from weak shales and limestones which formerly covered the Tuscarora arch.

The ridge shows no offsetting or displacement at the gap. The strongest evidence against important faulting is found in minor flanking ridges of Oriskany sandstone. Rocky Ridge on the northwest side and Stony Ridge on the southeast are not offset at the river. Because the Oriskany beds dip in opposite directions in each of these ridges significant transverse faulting would certainly offset one or both of the ridges.

FAULTS UNRELATED TO WATER GAPS

One aspect of the distribution of faults and water gaps supports the drainage history outlined by Johnson (1931). Up to now nothing has been said of faults which break the ridges but which have not been utilized by streams. If such faults are found, and especially if there are near-by water gaps across which the strata pass unbroken, then one may conclude that failure of the streams to occupy the fault lines is (1) an expected consequence of chance superposition from a coastal plain, but (2) incompatible with the hypothesis that during slow evolution of drainage from Permian time streams have sought out lines of weakness to form the water gaps. Five such cases were found.

(1) Three miles northeast of Tyrone, Bald Eagle Mountain shows a pronounced offset (Pl. 2). The ridge is part of the north limb of the great Nittany Arch and is upheld by the Tuscarora and Oswego formations which dip steeply northwest. The Tuscarora formation produces a prominent northwest ridge crest, while the Oswego formation produces a somewhat lower bench paralleling the Tuscarora ridge along its southwest side. The fault which produces the offset in Bald Eagle Mountain cuts the ridge obliquely and trends east. The apparent horizontal displacement is slightly more than a quarter of a mile. The fault has been described and mapped by Leslie (1892, p. 630) and Butts *et al.* (1939, p. 78). There is a col about 300 feet deep in the ridge crest where the fault crosses, showing that the fault is a line of weakness. At Tyrone, 3 miles southwest of the col is a water gap of the Little Juniata River. There is therefore in close proximity a transverse fault breaking a major ridge but not utilized by a stream, and a water gap at a place where no weakness is apparent. This does not fit a hypothesis in which Appalachian streams extended themselves northwestward across the folded belt by utilizing zones of weakness in the major ridges. On the other hand, the situation is expectable if the Little Juniata is a superposed consequent stream.

(2) In the southeast corner of the Tyrone, Pennsylvania, quadrangle Canoe Mountain is broken by a fault (Pl. 2). The ridge is composed of Tuscarora quartzite which dips northwest at a moderately high angle and forms the southeast limb of Scotch Valley syncline. Horizontal displacement of the fault is about 500 feet in the Tuscarora formation (Butts *et al.*, 1939). The ridge-crest line makes a jog at the fault, and there is a conspicuous saddle where the fault crosses the ridge. The saddle does not seem to be a wind gap, but merely a low col resulting from more rapid weathering and erosion of the fault line. Here then is a second case in which a recognizable transverse fault is not utilized for a water-gap site.

(3) Canoe and Lock mountains in the vicinity of the Frankstown Branch of the Juniata River would be a single continuous ridge of Tuscarora strata were it not for a fault (Pl. 2). The ridge trends a little east of north, while the fault strikes more nearly northeast; the offset ends overlap about a mile. Rogers (1858, p. 896) described this as a striking example of an oblique dislocation. The Frankstown Branch of the Juniata River takes a subsequent course between the offset ends of the ridge, but part way through the natural gap it turns sharply southeastward and flows across the end of Lock Mountain through a narrow water gap. The subsequent valley, however, continues northeastward between the overlapping ridge ends. At the water gap, however, no fault was found. Any postulated drain-

age history must reasonably explain the remarkable fact that the river takes this difficult course in preference to an easy path into the adjacent lowland.

Under the hypothesis of subsequent-stream development the river would occupy the entire subsequent lowland between the offset ends of the ridge. The fact that it does not weakens this hypothesis. Under the regional-superposition hypothesis the water gap across Lock Mountain would represent part of the original superposed stream, and the subsequent course between the offset ridge ends a later modification. Whether this superposition was regional or local, in the latter case from weaker formations in the fold, may be debatable.

(4) Stone Mountain, 10 miles northwest of Lewistown, is offset by an oblique fault so that the ends of the broken ridge overlap about a mile (Pl. 2). The structure, recognized by Rogers (1858, p. 896) and mapped by Leslie (1892, p. 628), is clearly expressed in the topography along the boundary between the Allensville and Lewistown quadrangles. The fault line is marked by a col about 200 feet lower than the ridge crests on either side, into which head branches of Standingstone and Kishacoquillas creeks. Two miles farther northeast an offset in Stone Mountain, causing an offset in the Huntingdon County-Mifflin County boundary line, and the topographic aspect of oblique benches on either side of the ridge suggest an oblique fault which appears to join with the one first described.

These faults break the resistant Tuscarora formation of Stone Mountain, yet they are not utilized as water-gap sites. The zone of weakness, although it has topographic expression, gives rise to cols only. Were one to interpret the cols as wind gaps abandoned by the streams which carved them, one would have to admit that the fault line is not a zone of major weakness.

(5) In northeastern Everett quadrangle, Tussey Mountain is complicated by a small southward-plunging syncline and anticline (Pl. 2). Along the axis of the syncline a fault causes the western part of the synclinal nose to be pushed farther north than the eastern counterpart. The headward portion of a small south-flowing subsequent stream has grown eastward across the fault line in weaker rock between the offset ends of the Tuscarora formation, but no true water gap has formed.

Five miles south of this area is the gap of Yellow Creek, a southeast-flowing stream draining the center of the Nittany anticline. At this gap no evidence of faulting was found, and none is suggested by the topography. Thus, a conspicuous fault line breaking the resistant strata has not been fully utilized, whereas near by a water gap cuts through an apparently unbroken portion of the same ridge.

The five cases described are the only instances of major transverse faulting which the writer has observed in the area studied. All five have been mapped by the Pennsylvania Geological Survey. Three are especially significant because of nearby water gaps at points where no weaknesses could be detected. Where easy passage through a ridge of resistant rock was available, it was not taken. This behavior scarcely fits the hypothesis of subsequent streams, which assumes stream development along lines of faulting. On the other hand a chance distribution of southeast-flowing streams with respect to the underlying structures is apparent, and this characteristic favors the regional-superposition hypothesis.

The faults, except possibly that northeast of Tyrone, are not truly transverse in

orientation but merely cut obliquely across the ridges and make an acute angle with the strike. They are more accurately described as strike or thrust faults. Three of the five are parts of a single, extensive thrust fault (Ashley, 1931). This type of fault cannot produce the aligned groups of gaps characteristic of central Folded Appalachian drainage.

The general absence of transverse faults is borne out by a review of the literature. The Second Pennsylvania Geological Survey atlases of the Northern, Eastern Middle, Western Middle, and Southern anthracite coal fields were examined. The coal basins are shown on Mine Sheets drawn to the scale of 1:9600 upon which are indicated the geology and mine workings, and occasionally structure contours on a coal bed. No suggestion of transverse faulting was observed on these maps. It seems unlikely that any appreciable faulting would have escaped detection, especially since the offsetting of coal seams along such faults would be of great economic importance. The more recent structure map of the Northern Anthracite coal basin (Darton, 1940) likewise shows no transverse faulting within that area. The various geologic maps and reports of the Pennsylvania Survey covering the remainder of the folded belt show that transverse faults, either tear faults produced at the time of folding, or later normal faults, are exceedingly rare. The few cases already mentioned were noted and described at an early date.

HYPOTHESIS OF MODIFIED CONSEQUENT DRAINAGE

STATEMENT OF THE HYPOTHESIS

Meyerhoff and Olmsted (1936) outlined a history of central Appalachian drainage in which major streams date from Permo-Triassic time and originated as consequent streams on folds and thrust sheets. Departing from Davis' (1909, p. 451–453) concept of a regional divide located initially in what is now the crystalline belt of the Older Appalachians, Meyerhoff and Olmsted (1936) postulate that the divide may have been located within the Folded Appalachian belt, and that the present seaward drainage in the area southeast of that early divide has been directly derived from initial Permian consequent streams. Because the divide today lies in the Appalachian Plateau province, a considerable drainage reversal must have taken place, although this matter is not mentioned.

This hypothesis will be referred to as the hypothesis of modified consequent drainage. Consequent streams flowed down the flanks of asymmetrical anticlines and the surfaces of thrust sheets or passed from one syncline to another across low places in the crests of intervening anticlines. Longitudinal consequents flowed down the axes of plunging synclines and joined the transverse consequents. Later modifications resulted from development of subsequent streams, either along weak-rock belts or in transverse fault lines. The initial consequent streams, while perhaps retaining their positions, came in some instances to bear anomalous relationships to the underlying structures as denudation caused a shifting of outcrops or the exposure of new structures beneath thrust sheets.

Certain discordant relationships between transverse streams and underlying structures are explained by types of superposition hitherto given little or no atten-

tion. Streams superposed from consequent courses upon the back slope of a great overturned fold are considered to have cut the abundant Kittatinny Mountain (Pl. 2) wind and water gaps. Other streams, within or southeast of the Great Valley, are thought to have been superposed from thrust sheets.

The various genetic classes into which the larger central Appalachian streams would fall according to the hypothesis of modified consequent drainage are: (1) streams consequent upon Permian topography, both longitudinal and transverse; (2) superposed streams, including those superposed from overturned folds and thrust sheets, as well as subsequent and consequent streams locally superposed upon unrelated structures beneath; and (3) subsequent streams, longitudinal in weak-rock belts, and transverse in fault zones.

POSITION OF THE PERMIAN DIVIDE

The principal point wherein the hypothesis of modified consequent drainage differs from Davis' hypothesis is the Permian location of the divide between Atlantic-slope streams and inland-flowing streams. Meyerhoff and Olmsted (1936, p. 25, 27) recognize that the core of ancient crystalline rocks along the Blue Ridge-Reading Prong-Green Mountain axis suggests that this belt represents the line of greatest elevation in Permian time, but they postulate that overturned folds and thrust sheets, since largely removed, caused the Permian divide to lie within the folded belt northwest of the Great Valley but not so far inland as the Allegheny Front. In Pennsylvania and New Jersey thrust sheets are considered to have extended across the Great Valley to Kittatinny Mountain but probably not beyond (Pl. 1). Thus it is implied that overturned folds carried the divide the remaining distance inland across the folded belt. Farther north, thrust sheets are postulated to have extended across the Hudson lowland.

Mackin (1938, p. 32) analyzed the possible effect of overturned folds and thrust sheets in shifting a divide in the direction of overturning and thrusting. He states that although overturning may result in the asymmetric location of the fold crest with respect to the two limbs, it does not necessarily affect the position of the master divide. The same is true of thrusting. Figure 4 illustrates the independence of individual folds and thrust sheets from the regional divide. In both A and B, the folds are overturned in the same direction. For individual anticlines the crest of the fold may form a local divide, asymmetrical with respect to the synclinal troughs on either side; but the master divide is determined by the topographic culmination of the series of folds, and not by direction of overturn. C and D indicate two hypothetical forms produced by low-angle thrusting, and it is assumed that the head of each thrust is the highest part of that sheet. Although in each case the slopes of an individual thrust sheet are highly asymmetrical, the regional divide lies on the culminating crest of the series as a whole. Thus the asymmetry of individual local structures does not necessarily indicate asymmetry of the regional divide with respect to the root of the mountain belt.

Mackin (1938, p. 31–32) discussed other doubtful aspects of divide location. The overthrusts considered to have in part determined the position of the divide are assumed to have extended northwestward over the Great Valley, although there

seems to be little evidence of this. It is further assumed that overturning and overthrusting would have had a cumulative effect in shifting the regional divide westward, but Mackin (1938, p. 31–32) has stated that the same stream cannot at once be consequent on both folded strata and a thrust plate that covers the folds.

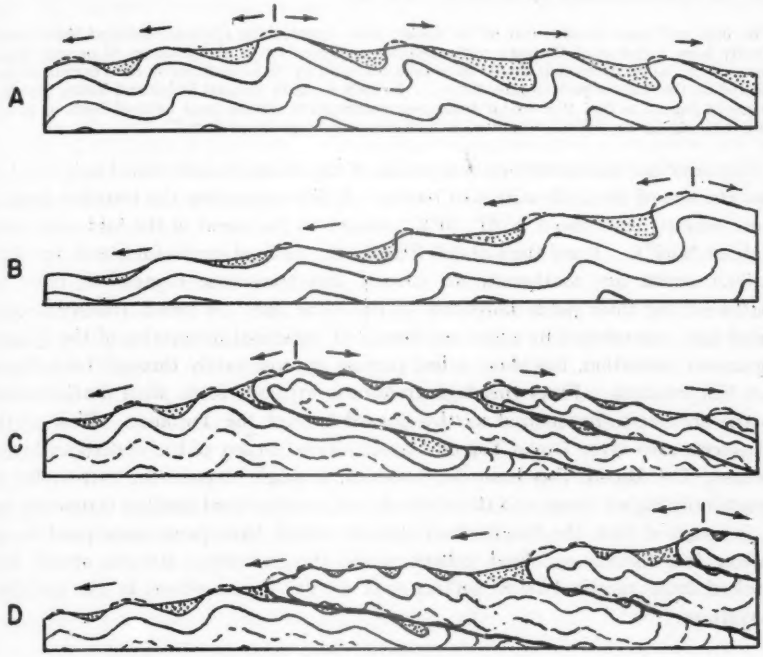

FIGURE 4.—*Hypothetical structure sections*
Showing independence between individual folds or thrust sheets and the position of the regional drainage divide.

An even more serious objection to the position of the divide is that it has not been placed far enough inland to remove entirely the vexing problem of drainage reversal. This problem may be dismissed only if the divide is placed at, or inland of, the Allegheny Front. Such a postulate, requiring the plateau margin to have initially been the highest part of the entire Appalachian mountain belt, conflicts with present geologic structures.

CONSEQUENT ORIGIN OF APPALACHIAN STREAMS

Particular emphasis has been placed on the Permian-consequent origin of major central Appalachian streams. Although drainage is considered to have been modified, Meyerhoff and Olmsted (1936, p. 29, 33) have cited certain instances in which present stream courses correspond with major structural features. Only two examples in the area here treated are mentioned, the first of which is the group of longitudinal streams flowing southwest from the Catskills, and considered to have

been derived from longitudinal Permian consequent streams. The second is a reference to "the natural movement of intermontane streams toward the orogenic depression in the vicinity of Harrisburg."

Mackin (1938, p. 42) has said of the relation between major streams of the folded belt and the structures upon which they flow:

> "The first and most obvious test of the theory that Appalachian streams inherited their courses directly from a system of Permian consequents is a comparison of the positions of present stream courses with the positions that should have been occupied by major streams on the Permian surface, restored on the basis of present structures. . . . In such a region (central Ridge and Valley province) we might expect to find that major transverse consequents should cross belts of folds in or near transverse sags in the folded structures, if such structural sags are present."

The principal transverse syncline or sag of the Pennsylvania folded belt (Pl. 1) is now the site of the anthracite coal basins. A line connecting the broadest parts of these basins trends about N.20°–30°E., oblique to the trend of the fold axes which is about N.65°E. From the Catskill Plateau the pitch of gentle folds and dip of the plateau strata are southwestward toward this transverse depression; from the southwest the folds pitch northeast to the same line. A broad transverse anticlinal axis, represented by numerous breached anticlinal mountains of the Silurain Tuscarora formation, lies along a line passing approximately through Lock Haven and Shippensburg. From this high transverse arch the folds pitch northeastward to the anthracite basins and southwestward toward the Potomac. West of this transverse arch is the Broad Top coal basin, similar to any of the eastern anthracite basins. The Broad Top basin is, however, a single depression, surrounded by structurally higher areas, and therefore cannot be considered another transverse sag. It is assumed that the longitudinal streams would have been consequent in the narrow longitudinal synclinal valleys, while the transverse streams would have resulted from synclinal lakes spilling over at the lowest places in the anticlinal mountains.

LONGITUDINAL STREAMS

The West and East Branches of the Delaware and the upper course of the North Branch of the Susquehanna flow southwestward from the Catskill Plateau toward the anthracite basins and thus follow the dip of the plateau beds and pitch of the shallow folds. These streams Ruedemann (1932) regarded as Permian consequents, and that interpretation is to some extent utilized by Meyerhoff and Olmsted (1936, p. 29) who recognize that these streams are now subsequents but suggest that they were derived from closely related Permian ancestors. Mackin (1938, p. 45) replied that, while such an origin is possible, the admission that the streams are now subsequent permits the interpretation that they have grown headward in weak-rock belts regardless of the origin and form of the pre-existing drainage. Since there is this satisfactory alternate explanation, these southeast-flowing streams do not necessarily support the hypothesis of modified consequent drainage.

The southwest-flowing portion of the Delaware River between Port Jervis and the Delaware Watergap (Pl. 1) also flows in the direction of regional pitch of folds, but here the structural relations are more obvious. The present stream is undoubt-

edly subsequent in origin since it occupies a valley eroded on weak Devonian beds in a monoclinal sequence. No strongly defined syncline appears to lie close to this stream. Instead, the expectable position for a major longitudinal consequent would be about 25 miles northwest, where the Pocono formation suggests a major trough axis. The present Delaware can scarcely be considered as directly derived from a consequent stream so far away.

More significant is the 60-mile stretch of the North Branch Susquehanna River where it flows southwest along the Wyoming coal basin between Pittston and Sunbury (Pl. 1). Throughout this distance the river flows exactly opposite to the direction of the pitch of the folds. The initial consequent must have flowed northeast along the axis of the Wyoming Syncline (Davis, 1909, p. 452). The present stream is thus clearly an independent subsequent stream not directly derived from a Permian consequent.

The West Branch Susquehanna River between Lock Haven and Williamsport (Pl. 1) flows 35 miles in a subsequent valley paralleling the Allegheny Front. This direction coincides with the pitch of the Nittany Anticline around the end of which the stream turns to flow southward. Although the direction of flow is correct for an ancestral Permian consequent, there is no strongly defined synclinal axis northwest of the Nittany Anticline, the northwesternmost major fold of the Ridge and Valley province. Minor folds in the plateau to the northwest suggest that a Permian stream could have flowed parallel with the present one, but at some distance northwest. With excavation of the subsequent valley that consequent stream may have been diverted into the new subsequent course. This part of the West Branch Susquehanna can therefore be explained under the hypothesis of modified consequent drainage but does not demonstrate its correctness.

The Juniata River between Mt. Union and Mifflintown flows northeastward parallel with the strike of the folds for about 25 miles (Pl. 1). Here it is difficult to determine the directions of pitch of structures under the stream, which lies in a synclinal area between two anticlinal ridges of the Tuscarora formation. For a few miles northeast of Mt. Union the pitch is definitely southwest. Near Lewistown, the Juniata passes between the overlapping noses of two anticlines which pitch in opposite directions. Thus, no direction of flow can be assigned to an ancestral Permian consequent in this region. Possibly the Juniata here does flow in the same direction and on approximately the same line as a Permian consequent, or it may be purely subsequent in origin and unrelated to pre-existing drainage.

The Raystown Branch of the Juniata River between Everett and Huntingdon flows northeastward for about 35 miles along the northwest margin of the Broad Top coal basin, parallel with the strike of the beds (Pl. 1). The southwestern third of this longitudinal stream flows with the pitch of the syncline, but the other two-thirds flows up the pitch. Thus for most of its course the stream flows in the opposite direction from that expected of a Permian consequent. However, direction of flow may have initially been the same as the stream now follows—toward the northeast,—because the Broad Top basin would at first have been closed and would have had to drain. Davis (1909, p. 453) believed it drained northeastward toward the eastern Anthracite basins, and from an inspection of the geologic map this seems plausible.

In any case the present stream is clearly subsequent because it lies on the common flank of the Nittany anticline and the Broad Top basin. Davis described in detail how initial consequent drainage was here transformed into a subsequent system.

All streams considered, except the West Branch Susquehanna between Pittston and Sunbury, could have derived from Permian consequent streams. They do not, however, represent actual Permian streams but are subsequents which grew in adjacent weak-rock belts and diverted earlier streams. Longitudinal subsequent streams must be expected to form under a great variety of postulated drainage histories. They cannot, therefore, be of critical value in estimating the validity of a theory. A possible exception would be major streams flowing down the pitch of synclines with such uniformity as to suggest that longitudinal Permian consequents had escaped diversion through all succeeding erosion cycles. Such evidence of Permian drainage was not found.

TRANSVERSE STREAMS

If the major drainage of the central Folded Appalachians is in part inherited from a consequent Permian system, some of the transverse streams might reasonably be expected to occupy low places in the crests of folds which they cross. Meyerhoff and Olmsted (1936, p. 29, 31) state that the Permian divide lay northwest of Kittatinny Mountain, but southeast of the Allegheny Front. This location resulted from thrusting and overturning of folds. It is difficult to believe that these structures could have caused the divide to be pushed so far northwest of the limits of major overturning and overthrusting, but unless the divide is postulated to lie near the Plateau margin, the hypothesis is confronted with the same difficulties as was that of Davis (1909, p. 459–461)—namely, that a later drainage reversal is required. Because no reversal is mentioned, it is assumed that the initial divide lay close to the Plateau margin. Only on this assumption could most of the transverse southeast-flowing streams have descended directly from consequent streams which found their way across sags in the folds.

The southeast-flowing part of the Delaware River above Port Jervis (Pl. 1) lies midway between the highest structural elevation of the Catskill Plateau and the low transverse axis of the anthracite basins. It cannot therefore be considered as consequent on any initial transverse sag.

The southeast-flowing Towanda-Pittston segment of the North Branch Susquehanna River is not coincident with any transverse structural sag but lies 10 to 20 miles northeast of a chain of shallow Pottsville basins which would be the expectable site of a Permian consequent stream. At Pittston the North Branch does reach the center of the Wyoming syncline and thus intersects the major transverse structural depression marked by the bottoms of the eastern anthracite coal basins (Pl. 1). The river, however, does not flow parallel with that axis but intersects it at approximately a 70-degree angle. As this stream segment lies within the Appalachian Plateaus, it must, according to the hypothesis under examination, either be a reversed consequent or a new stream developed by headward growth because it lies inland from the postulated Permian divide. Whichever explanation is chosen, the stream cannot be considered related to a Permian consequent.

The Lehigh River, from its origin in the Pocono Plateau, flows south and southeast across the various folds to the Reading Prong (Pl. 1). This part of the Lehigh River does not occupy a transverse structural sag but is located east of a line connecting the deepest parts of the anthracite coal basins. The Schuylkill, however, does originate approximately in the transverse depression, and Davis (1909, p. 463) considered this stream a descendant of the reversed Anthracite River. This portion of the Schuylkill can therefore reasonably be explained by the hypothesis of reversed consequent drainage.

The Susquehanna River above Harrisburg flows across two synclines and seems to lie about midway between the transverse axis of highest elevation on the west and the anthracite basins axis on the east (Pl. 1). In this respect the river does not conform to any structural sag and therefore cannot be regarded as of Permian consequent origin. For this reason the reference by Meyerhoff and Olmsted (1936, p. 33) to "the natural movement of intermontane streams toward the orogenic depression in the vicinity of Harrisburg" is difficult to reconcile with field evidence. Elsewhere they explain this part of the Susquehanna as a subsequent stream located, as Ashley (1935, p. 1406) has suggested, on a transverse fault breaking the resistant ridge-forming strata.

Other major southeast-flowing streams are the West Branch Susquehanna from below Williamsport to Sunbury, and above Lock Haven, and the Little Juniata-Juniata between Tyrone and Mt. Union. It is clear (Pl. 1) that none of these streams is located in a transverse structural sag. They cannot therefore be considered as Permian consequents.

SUPERPOSITION FROM ASYMMETRIC FOLDS

Meyerhoff and Olmsted (1936, p. 33–35) believe that the numerous wind and water gaps in Kittatinny Mountain are not normal results of erosion of open folds. As an alternative to the former presence of a coastal-plain cover over the ridge, they suggest that streams flowing down the back slope of a great overturned anticlinal fold became superposed across ridges in the underfolded northwest limb (Meyerhoff and Olmsted, 1936, p. 26, 34). To illustrate the process of superposition they have prepared a series of four diagrams (Fig. 5). Diagram B shows an asymmetrical anticline. Streams flowing down the gentle back slope are several times as long as those flowing down the steep forward slope. The authors state that the much greater volume of the longer streams would delay migration of the divide toward the geographical center of the fold, thereby permitting those longer streams to cut through the fold and produce the water gaps and wind gaps shown in diagrams C and D.

They further report numerous and almost equidistantly spaced gaps in portions of Kittatinny Mountain along which the beds are overturned, whereas along those parts of the ridge where dips are steep to the northwest gaps are absent. The correlation is considered too strong to be the result of chance superposition. The presence of relatively few wind and water gaps in ridges northwest of Kittatinny Mountain they regard as support for the hypothesis, because in that area folds are open, and the requisite overturning absent.

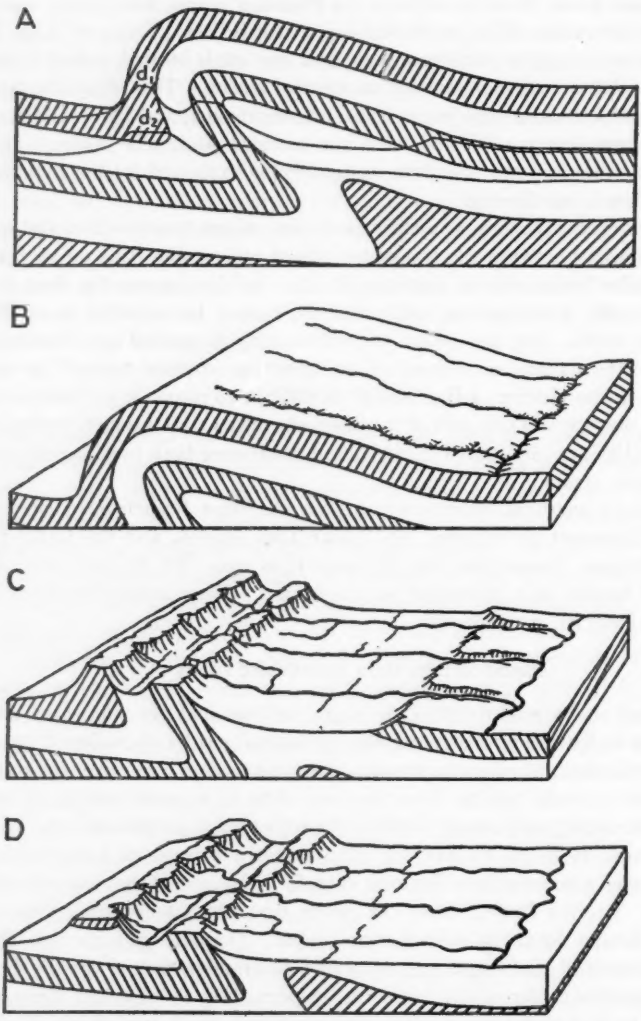

FIGURE 5.—*Diagrams showing stream superposition in asymmetric folds*
Suggests a possible explanation of wind- and watergaps in Kittatinny Mountain (Meyerhoff and Olmsted, 1936).

These diagrams (Fig. 5) are based on the doubtful assumption that surface form and structure reproduce the contours of overturned folds in depth. The initial constructional surface shown in diagram B assumes that the fold when completed was essentially unmodified by erosion, whereas it seems likely that surface form must have been profoundly altered during the process of folding. The overturned limb has a slope of about 75°, but it is doubtful that mass movements and stream

action would have permitted such a steep slope to persist throughout folding. That the uppermost strata of the fold would bend in conformity with those at depth seems unlikely. If the rocks were sufficiently indurated to resist erosion to the degree indicated, one might expect them to be fractured rather than folded. Thus the initial form postulated in these diagrams seems in several respects improbable.

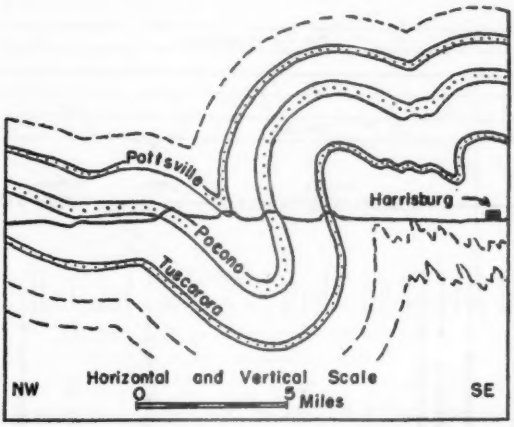

FIGURE 6.—*Restored section of the region northwest of Harrisburg*
After Leslie (1892)

Effectiveness of the superposition represented by the diagrams must depend on the streams maintaining their positions and lengths throughout a long period of denudation. The longer streams would have, at the point where they join the longitudinal trunk stream at the base of the fold, a greater volume than the shorter opposed streams where they join a longitudinal stream; but this fact has no immediate bearing on the stability of the divide. Near the divide opposed streams would have essentially similar volumes, whereas their gradients would be strongly contrasted. The shorter, steeper streams would cut quickly into the slope, and the divide would migrate rapidly toward the geographical center of the fold, thereby preventing superposition.

More or less equidistant spacing of initial consequent streams might result if the fold were as smooth as diagram B indicates; but to consider that they could maintain a balance from stage B to C seems a departure from the normal drainage history of breached anticlines in folded mountains. Only a small proportion of the streams breaching the crest of a young anticlinal mountain commonly survive. The others are diverted by subsequent tributaries of more powerful neighbors, and such integration goes on throughout denudation and results in fewer water gaps through the flanking ridges. The diagrams show stream integration beginning between stages C and D, but one must inquire why this did not occur in the long erosion history between stages B and C.

One may also question whether the fold illustrated in A is a reasonable reconstruction of the fold of which Kittatinny Mountain is a part. Figure 6 is a restored

section of the region north of Harrisburg (Leslie, 1892, p. 277). Although the reconstruction is largely interpretation, it is appropriate to the structures now exposed. The initial divide lies south of Kittatinny Mountain, almost at Harrisburg, with a minor divide in the latitude of Kittatinny Mountain. Under these conditions it is not evident how the postulated superposition occurred. The initial

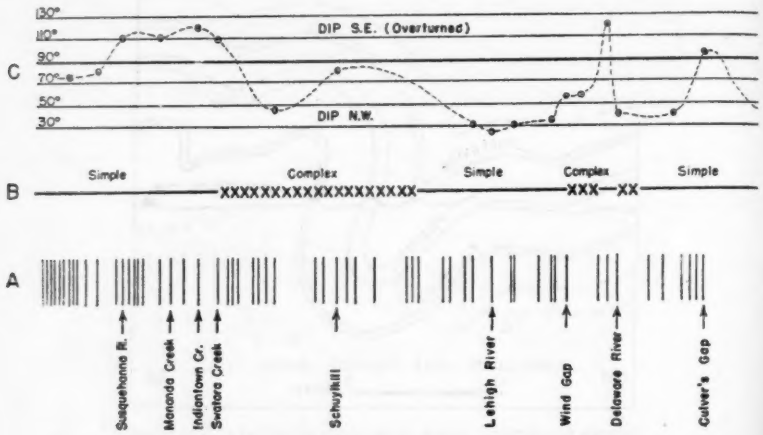

FIGURE 7.—*Diagram of the relations between wind and water gaps and the structure of Kittatinny Mountain*

surface shown could not have been completed without considerable modification by destructional forces; but it shows that an impartial interpretation of the structure is unfavorable to this phase of the hypothesis. Chamberlin's (1910, p. 233) restoration of this structure has similar characteristics. The hypothesis of superposition from an overturned fold thus seems to rest on a precarious foundation.

To test their hypothesis, Meyerhoff and Olmsted (1936, p. 34) have considered the relation of wind and water gaps of Kittatinny Mountain to the degree of overturning. They state:

"It was found that the formational structures are consistently and invariably overturned along those sections of the ridge in which wind- and watergaps are numerous, and almost equidistant. Where, on the other hand, the gaps are absent, the dips are steep to the northwest and there is no overturn."

To indicate the relation of dip of Kittatinny Mountain beds to the presence of gaps, the writer constructed a diagram (Fig. 7). The short, vertical lines (A) indicate the positions of wind and water gaps as shown on topographic maps and listed by Ver Steeg (1930) and are spaced proportionately to actual distances along Kittatinny Mountain. Part (C) is a curve showing maximum dips of the ridge-forming strata for corresponding points along the ridge. Near the Susquehanna water gap are five wind gaps, close together and nearly equidistant from one another, and this part of the ridge is overturned (above the line labeled 90°). A few miles west of the Susquehanna gap are nine wind gaps, even more closely spaced and nearly equi-

distant. Here the ridge-forming strata are not overturned but dip steeply to the northwest. This conflicts with the general statement by Meyerhoff and Olmsted. For a considerable part of the distance between Swatara Creek and Pen Argyl Wind Gap the strata of Kittatinny Mountain dip 30°–50°NW. Along this stretch wind and water gaps are farther apart than those near the Susquehanna gap but

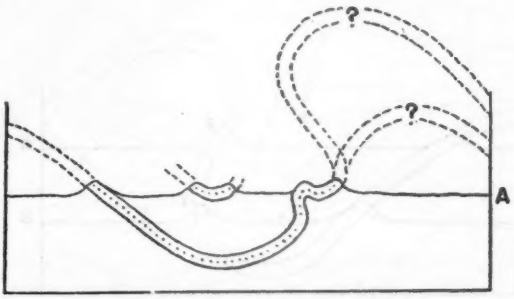

FIGURE 8.—*Minor flexure on the limb of a fold*
Producing a broad ridge of complex structure

are nevertheless numerous. Under the hypothesis being examined, these gaps should not be present. Two miles west of the Delaware Watergap, Behre (1927, p. 86) has found dips of 55°–80°S., indicating local overturning. Although Tott Gap is located at this point of overturning, the beds are not overturned at the adjacent Fox Gap and Delaware Watergap. Near Culver's Gap the strata are again overturned; a group of four wind gaps, closely and equidistantly spaced, are located somewhat southwest of the overturned portion of the ridge. Moreover, gaps are almost as abundant in many other places where there is no overturn. Thus examination of the gaps and structure of Kittatinny Mountain fails to reveal the correlation postulated by Meyerhoff and Olmsted.

The presence of minor folds and flexures in the strata of Kittatinny Mountain is further complication. The solid black line, Fig. 7, B, represents relatively simple or monoclinal structure, while the X's indicate minor folds striking nearly parallel with the regional trend. Near the Susquehanna River the structure is relatively simple, but between Swatara Creek and a point about midway between the Schuylkill and Lehigh Rivers a minor fold broadens the ridge. Again, between Wind Gap and the Delaware Watergap two small folds produce the Big and Little Offsets. For a few miles northeast of the Delaware Watergap another flexure complicates the ridge. How these smaller structural features may upset a simple correlation of dips with gaps is indicated (Fig. 8). Although the fold as a whole is overturned, exposures at the erosion level indicated may give low-dip readings which would suggest no overturning. Flexures along a substantial part of Kittatinny Mountain make hazardous any correlation of wind and water gaps with dips of the ridge-forming strata.

Of greater importance than the field facts is a consideration of the validity of the test itself. Dips now observed in Kittatinny Mountain cannot be taken as a safe

indication of the presence or absence of overturning. An overturned fold (Fig. 9) is beveled at two erosion levels. When the region is at the level *A*, strata in the ridge *X* will be overturned, showing that the fold is overturned. At the lower level, *B*, strata at *Y* will not be overturned, and there may be no means of determining the structure of the eroded portion. For this reason, proof of superposition from an overturned fold cannot safely be based on a simple correlation of dips with gaps.

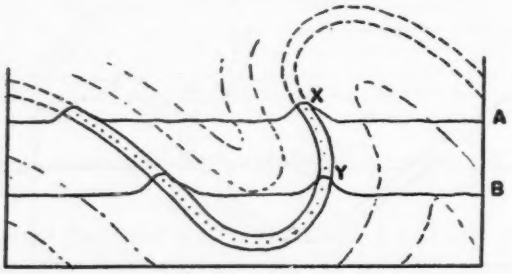

FIGURE 9.—*Overturned fold beveled at two successive erosion levels*

SUPERPOSITION FROM THRUST SHEETS

Meyerhoff and Olmsted (1936, p. 35) have suggested that some southeast-flowing streams of the central Folded Appalachians originated as consequents on thrust sheets. Influence of the thrusts is considered to have extended across the Great Valley in Pennsylvania, perhaps as far northwest as Kittatinny Mountain. Farther north, the Hudson-Champlain Valley may also have been overridden by thrust sheets. It is postulated that consequents flowed from the forward margins of the thrust sheets eastward or southeastward down the gentle back slopes. Denudation would have caused considerable recession of the low-angle thrust sheets, and some consequent streams may have become superposed upon folded structures beneath.

Meyerhoff and Olmsted (1936, p. 28) have illustrated evolution of a thrust sheet (Fig. 10). These diagrams are intended to suggest an interpretation of the Catskill-Hudson Lowland-New England relationships. In the second diagram the sheet has receded many miles, and although subsequent drainage is now dominant, one of the initial consequents still flows across a synclinal mountain.

The hypothesis requires that thrust sheets which now bound the Hudson-Champlain lowland on the east were produced in the Appalachian orogeny. If, however, the thrusting is much earlier, persistence of streams throughout later orogenies would be most unlikely. Keith (1923, p. 105–106, 137) considers the Green Mountain thrusts to be Permian on the basis of the physical similarity between thrust faults in the Green Mountains and those of the Southern Appalachians. Supporting stratigraphic evidence is lacking, and, moreover, similarity of structures hardly provides satisfactory evidence of simultaneous deformation. Schuchert (1930, p. 717) states that the intensive folding and thrusting of the Taconic and Green mountains occurred in the Taconian (Ordovician) orogeny as proved by unconformable superposition of Silurian strata on deformed Ordovician rocks of

northeastern Pennsylvania, northern New Jersey, southeastern New York, northern New Hampshire, and localities farther north. Clark (1934, p. 13, 18–19) states that all thrusts west of the Oak Hill slice seem to have occurred during the Taconian orogeny and offers as evidence the unconformable Silurian strata upon deformed

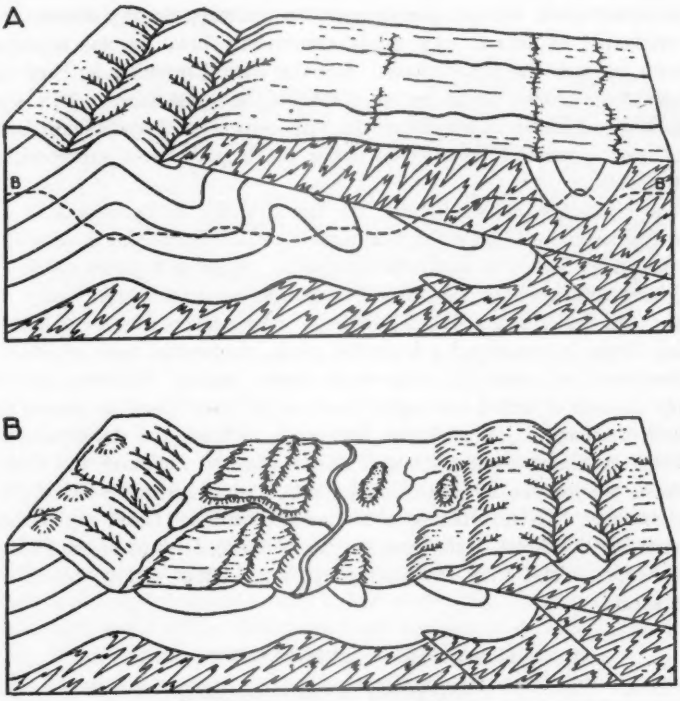

FIGURE 10.—*Hypothetical diagrams showing recession of a thrust sheet*
Suggests an interpretation of the Catskill-Hudson Lowland-New England relationship (Meyerhoff and Olmsted, 1936).

Cambro-Ordovician beds. Kay (1937, p. 287–288) agrees with the assignment of deformation in eastern New York, Vermont, and southern Quebec to the Ordovician Taconian Orogeny.

Thus the principal thrusts along the western border of the Taconic and Green mountains, to which Meyerhoff and Olmsted are apparently referring, seem to have been produced in the late Ordovician, rather than in the Permian Appalachian orogeny. Southeast-flowing streams originating on the thrust sheets must therefore have persisted since the close of the Ordovician. Such a history is scarcely plausible; the Acadian orogeny of the Devonian seriously affected New England and was the source of Catskill delta sediments transported westward across the area of Taconian thrusting. Thus the drainage on the thrusts, regardless of its initial direction, was westward in the Devonian.

The first diagram (Fig. 10) is apparently not intended to represent the initial topography following thrusting, because the forward edge of the thrust sheet includes none of the corresponding strata beneath the fault plane. Not only does this feature suggest considerable erosion, but the upper surface of the thrust sheet seems to be a peneplane, beveling the complex crystalline structures and an infolded syncline. On the thrust block are two parallel east- or southeast-flowing streams. Having once originated as shown, they might conceivably have become superposed on structures beneath the thrust plane. But the critical question is: How did these streams acquire courses across the initial folds of the thrust sheet? This is precisely the question which is under discussion with respect to Appalachian drainage in general, but it remains unanswered under the hypothesis of superposition from thrust sheets.

In the crystalline belts adjoining the Great Valley in Pennsylvania and New Jersey the existence of thrust sheets and their possible former northwestward extent over the Great Valley is relatively uncertain. Stose and Jonas (1935) and Kay (1941) have considered the possibility of thrust sheets over the Great Valley. Meyerhoff and Olmsted (1936, p. 27–28) follow a more conservative view in which the Reading Prong is considered a fractured block, the several parts of which moved northwestward on relatively steep-angle thrust planes. Whether the structure actually consists of folded low-angle thrusts or of tilted blocklike masses the effect on initial drainage would probably have been to block the development of long transverse southeast-flowing streams rather than to facilitate it. Even if the hypothesis of superposition from thrust sheets is in all respects valid and a reasonable possibility, it is nevertheless of limited application. Other explanations must be sought for the direct, transverse southeast-flowing streams of the folded belt of Pennsylvania, inland from the farthest probable limits of any known extensive thrust sheets.

HYPOTHESIS OF PROGRESSIVE PIRACY

STATEMENT OF THE HYPOTHESIS

Thompson (1936, p. 1833–1836; 1939, p. 1323–1355) outlined a different history of Appalachian drainage from Permian time. His hypothesis of progressive piracy and local superposition embodies the following points: (1) The regional drainage divide, following the Appalachian Revolution, lay along the Blue Ridge-Reading Prong crystalline axis. (2) In the vicinity of the Potomac River and presumably also farther north, the divide has shifted across the entire folded belt. (3) The divide migrated because it was much closer to the Atlantic Ocean than to the Gulf of Mexico. Repeated eastward tilting has shortened and steepened Atlantic-slope streams and renewed their advantage over inland-flowing streams. Because tilting increases northward, shifting of the divide has been greater toward the north. (4) The divide migrates by slow normal headward erosion and piracy. Resistant ridges have impeded shifting of the divide but have nevertheless been breached by headward erosion, and the subsequent valleys beyond have been captured. (5) Water gaps have been produced where the ridge-forming strata are relatively thin or weak.

The Potomac and James rivers and the Hudson Gorge provide examples of this. (6) Other water gaps may have been produced by local superposition as resistant beds deeper in the folds were exposed.

The hypothesis is similar to one proposed by Davis (1909) in that the initial drainage was northwestward across the folds from a divide along the present belt of crystalline rocks. The important difference is that instead of reversal of major

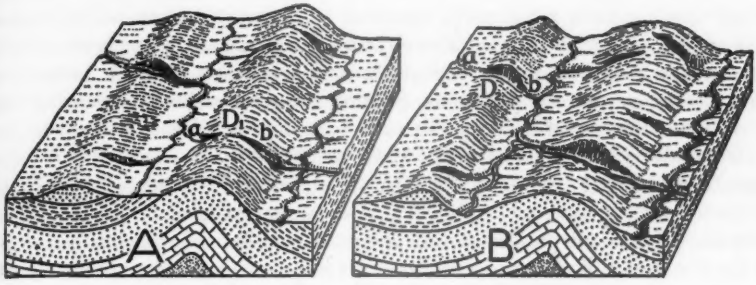

FIGURE 11.—*Diagram showing stream capture and migration of a divide*
(Thompson, 1939)

transverse rivers in their valleys, Thompson suggests that the transverse southeast-flowing streams grew without regard to the pre-existing transverse streams, giving rise not merely to a reversed system, but to new locations of many transverse elements. Both hypotheses differ from that of modified consequent drainage.

Thompson has deduced that under the hypothesis of progressive piracy water gaps should have been produced where the strata are thin or the rock weaker, but no emphasis is placed on possible control by transverse fault zones in the folded belt. It may have been assumed that transverse faults are absent or of little consequence, a view fully in accord with the results of the writer's investigations.

Although Thompson has intended the principles of the hypothesis of progressive piracy and local superposition to be applied to the Pennsylvania folded Appalachians, his discussion of specific streams has been confined to the southern Appalachians, the Potomac system, and the Hudson Gorge. The writer will deal with the principles involved and their application in the folded belt of Pennsylvania. While the discussion of principles may apply to both the northern and southern areas, the evidence obtained in one region does not necessarily support or invalidate application of the principles to the other.

FORMATION OF WATER GAPS BY HEADWARD GROWTH OF STREAMS

Thompson has described how water gaps may be formed by headward growth of streams shifting the divide inland across a folded belt. In diagram *A* (Fig. 11) the divide lies along the crest of an anticlinal ridge flanked by weak-rock lowlands in which graded streams flow at elevations differing by several hundred feet. Small tributaries *a* and *b* on either side of the ridge head opposite each other. In diagram *B* a water gap has been cut through the ridge, and the divide has been pushed back

to the next ridge. Unfortunately, critical intermediate stages are omitted, and doubtful phases of the process left undiscussed.

A questionable element of this hypothesis is an assumption that the small stream, *b*, will push the divide entirely across a high and broad resistant rock barrier and into the lower weak-rock lowland beyond. As discussed earlier, a divide is not acted upon by the master streams lying on either side of it, but by local weathering, mass movements, and unconcentrated runoff. Since these factors affect both slopes of the divide more or less equally, the latter will normally remain on the resistant rock so long as that rock stands above the adjacent lowlands, whatever may be the difference in their elevations. Meanwhile divides in the weak-rock lowlands will shift rapidly as subsequent streams, growing along the strike of weak-rock belts, work around the ends of resistant structures to capture and deepen lowlands beyond.

Only under special circumstances does it seem likely that a water gap will be produced in the manner assumed. The essential requirement is a difference in elevation of the two main streams so very great that of two opposing minor tributaries, each with steep headward profiles of similar gradient within the resistant barrier, one of the tributaries discharges so far below its neighbor that it can ultimately work clear through the obstacle and still have the steep upper portion of its profile lower than the trunk of the large stream flowing on the weak rock beyond. The difference in elevation between the two stream systems is then so extreme that a veritable escarpment exists between them, the ridge itself thus becoming a minor element of the topography, and the necessary conditions may exist for the divide to be shifted across the resistant rock belt.

No evidence of these conditions has been found in the Pennsylvania folded Appalachians. Where water gaps occur we find no traces of high-level lowlands on one side of the ridge and low-level surfaces on the other. Repeated instances of long, roundabout subsequent streams, flowing many miles along weak-rock belts to drain lowlands only a few miles back of resistant barriers testify to the comparative impotence of headward erosion in water-gap formation. The abundance of wind gaps proves that the process long operating in the Pennsylvania folded Appalachians is the opposite of that postulated for the hypothesis of progressive piracy. Water gaps are being abandoned rather than created through the normal development of longitudinal subsequent drainage.

Thompson (1939, p. 1334–1342) finds in the southern folded Appalachians some confirmation of his hypothesis in the occurrence of water gaps where ridge-forming strata are thinner or the rock locally weaker. While these factors may localize water gaps assuming the extreme conditions required for their formation were present—the evidence cited does not necessarily prove the validity of this hypothesis. Under any hypothesis water gaps will persist longer where resistant rock belts are thinnest or weakest.

EVIDENCE OF PROGRESSIVE PIRACY IN THE PENNSYLVANIA FOLDED BELT

To test the hypothesis of progressive piracy and local superposition the resulting drainage pattern should be deduced and compared with Pennsylvania drainage. Because divides migrate most easily along weak rock belts, the encroaching streams would avoid barriers of resistant rock insofar as possible. Although local cases of

superposition across pitching anticlines would occur, subsequent drainage would be developed to its highest degree because rock resistance would dominate growth of the stream system. Instead, the major phenomenon of Pennsylvania drainage is long segments of southeast-flowing streams, repeatedly crossing many resistant ridges.

The series of Susquehanna River gaps north of Harrisburg (Pl. 2), present despite an easy path for the river on weak-rock lowlands to the west, is perhaps the greatest single objection to the hypothesis. Drainage of the Little Juniata from Tyrone southeastward across the Nittany anticline instead of northeastward along a subsequent valley to the Susquehanna River at Lock Haven is also unexpected. Dual gaps, maintained independently but close together in a ridge just above the streams' junction, as in Tussey Mountain or at Pottsville and Tremont, are not reasonable consequences of subsequent-stream development. Abundant wind gaps in Kittatinny Mountain and in ridges north of Harrisburg, where only a few passages through the ridges would have sufficed, constitute still another unexplained group of features. The straightness and consistent southeast orientation of the several transverse stream elements, regardless of variations in strike of the folds (Pl. 1), likewise finds no explanation in the hypothesis of progressive piracy and local superposition.

HYPOTHESIS OF REGIONAL SUPERPOSITION

Objections raised to the hypothesis of regional superposition of Appalachian drainage are examined here to determine if valid points of weakness have been found in the hypothesis since its statement by Johnson (1931).

Meyerhoff and Olmsted (1936, p. 29) state:

"The adherents of the theory of regional superposition dismiss Appalachian structure as the initiating factor in localizing modern stream courses and as a determinant of the present geomorphic pattern, although in doing so they must ignore many structural and stratigraphic facts of fundamental importance."

Neither the inventor nor adherents to the hypothesis of regional superposition have ignored the structural and stratigraphic facts of the Appalachians, but Appalachian rivers ignore them to a remarkable degree. It was precisely because studies of Appalachian structures alone could not explain major features of Appalachian drainage that a new interpretation was sought. Consequent southeast drainage on a coastal plain must have been initiated without regard for structures in the deeply buried rocks. This, however, does not mean that structures have not profoundly influenced later stages of drainage development once the coastal-plain cover was removed. Full attention has been given to the influence of structure and stratigraphy on Appalachian drainage evolution by showing how they caused great development of subsequent streams and brought about extensive dismemberment of the initial consequent system (Johnson, 1931, p. 34–35, 64).

Meyerhoff and Olmsted (1936, p. 32) state:

"In the explanation of Appalachian drainage proposed in this paper, the consistent utilization of structural elements by all major streams is logical and necessary; but to fit the facts into a theory of regional superposition from a coastal plain, one must attribute to the consequent drainage on the marine cover such an uncanny insight into the foibles of the buried structures that the theory becomes incredible."

Earlier in the present paper the writer has shown that the structural elements postulated by Meyerhoff and Olmsted to have localized the major southeast-flowing streams in Permian time are not present in the folded belt. Streams now cross the belt oblique to the one major transverse synclinal axis and rarely occupy structural sags (Pl. 1). These conditions suggest a wholly fortuitous distribution of transverse drainage elements with respect to structure.

Meyerhoff and Olmsted state (1936, p. 32):

"Other tribulations imposed upon its adherents by the theory of regional superposition are found in the elaborate accounting necessary when wind- and watergaps are missing from a scheme that calls for aligned gaps, and when its proponents are compelled to postulate the dismemberment of large trunk streams which, if intrenched from a horizontal sedimentary cover upon buried ridges, must have possessed erosive powers adequate for permanent entrenchment throughout their courses."

The absence of many sets of wind gaps is a normal consequence of superposition on the ancient and now largely removed Fall-Zone peneplane, with extensive obliteration of wind gaps during the Schooley cycle. Dismemberment of transverse trunk streams by growth of longitudinal subsequents, far from being a doubtful process, has long been recognized as a major phenomenon of Appalachian history. Any interpretation which accounts for the present pattern of transverse streams and the distribution of wind gaps must recognize the operation of this process.

Regarding gaps of Kittatinny Mountain, which are said to have been superposed from an overturned fold, it is stated (Meyerhoff and Olmsted, 1936, p. 34):

"The geographic correlation between dips and gaps is so close that no plea of accident or coincidence can reconcile the actual situation to Barrell's hypothesis of marine overlap or to Johnson's assumption of a more widespread marine cover."

The writer shows earlier in the present paper that the supposed correlation of dips and gaps does not exist and that the assumed significance of such correlation rests on a misunderstanding of the relation of present dips to structures partially eroded.

Meyerhoff and Olmsted (1936, p. 34–35) state still another objection:

"For in the narrower ridges northwest of Kittatinny, there is scarcely a windgap or a watergap to match the many breaks of fluvial origin which cut through the broader barrier. . . . Were the streams entrenched from a coastal plain, . . . the gaps should be at least as numerous in the narrow barriers as they are in Kittatinny Mountain."

Many wind and water gaps are cut through the ridges northwest of Kittatinny Mountain. The writer counted those on topographic maps of the folded Appalachians within Pennsylvania. This area covers 78 quadrangles, and the 77 now mapped were included in the count. In Kittatinny Mountain there are 8 water gaps and 30 wind gaps, all recognized by Ver Steeg (1930). In the rest of the area, including only ridges of the Tuscarora, Pocono, and Pottsville formations, 201 water gaps and 36 wind gaps were counted. If we consider water gaps alone, we find 25 times as many in ridges exclusive of Kittatinny Mountain as in that ridge itself. If allowance be made for the many more miles of ridges in the folded belt than in Kittatinny Mountain alone, the distribution of water gaps is about the same throughout the region.

Although there are more than enough wind gaps elsewhere in the folded belt to match those in Kittatinny Mountain, wind gaps in the latter are more closely spaced.

Consideration of the process of wind-gap formation leads to the conclusion that this is an expectable result of regional superposition. Kittatinny Mountain is unique in that it is a continuous barrier from the Catskills to southern Pennsylvania. All streams flowing southeastward from within the folded belt must cross it. Thus in the contest for control of the weak-rock lowland immediately northwest of the ridge no stream would have a marked advantage over others by virtue of difference of rock resistance in its course. The ability of one transverse stream to send out a subsequent tributary and capture another above Kittatinny Mountain would depend chiefly on greater size. Farther inland, however, ridges of the folded belt are not continuous but end in anticlinal or synclinal noses (Pl. 1). Thus some streams must have been stranded across many more resistant rock barriers than others, making relatively easy the dismemberment of those less fortunately situated. The result of these factors would be that during the Schooley cycle wind gaps would have been produced rapidly within the folded belt but more slowly along Kittatinny Mountain. By the close of the Schooley cycle the earliest formed wind gaps would have been obliterated through extensive erosion toward the Schooley base level. Diversion of streams crossing Kittatinny Mountain was in many cases delayed until the Harrisburg cycle. This accounts for the greater abundance of wind gaps preserved in that ridge.

Ashley's (1935, p. 1406–1407; 1939, p. 9–10) objections to the theory of regional superposition seem to have arisen from his belief that many gaps of the folded belt coincide with fault zones. He regards the degree of coincidence to be greater than expectable under superposition. In an earlier discussion of this relation the writer has concluded that these objections cannot be sustained. Not only is the statistical analysis of gaps in relation to faulting unconvincing, but the critical major water gaps are free of significant transverse faults, whereas major transverse faults were not the loci of water gaps. The facts thus favor chance superposition from a coastal plain.

Thompson has cited a number of weaknesses in the hypothesis of regional superposition. A possible shortcoming in his argument against the older theory is that much of the unfavorable evidence is cited from the southern Appalachians, rather than from the northern portion to which Johnson applied the theory. By showing that a theory does not work in a region where it was not intended to apply one cannot invalidate it for another area where it effectively explains the facts observed.

Thompson (1939, p. 1329) states:

"As objections to the reversal of northwest consequent drainage, Johnson gives (1) the directness of the course, (2) the great length, and (3) the parallelism of the southeast-flowing streams. These objections may be open to question."

He then shows that the Potomac, James, and Roanoke rivers are nearly twice as long as the airline distance from their heads to the sea. Moreover, he finds that their courses possess neither great comparative length nor parallelism. Thompson may have misinterpreted Johnson's text. The latter wrote that southeast courses of the streams, not the entire streams themselves, are characterized by directness, great length, and parallelism (Johnson, 1931, p. 33). Between points marking the approximate ends of each southeast-flowing stream segment the streams only rarely

depart more than 5 miles from a connecting straight line. Directness is therefore a true characteristic.

The great length of the southeast stream courses is seen by comparison with remaining parts of the whole streams. With the exception of an unusually long stretch of the West Branch Susquehanna River between Pittston and Sunbury, longitudinal drainage elements connecting the transverse segments are consistently the shorter. They are not over 40 miles long, whereas five of the transverse stream sections exceed that length and range up to 70 miles (Pl. 1).

Approximate parallelism is a striking characteristic of the southeast-flowing stream segments (Pl. 1). Absolute parallelism is not to be expected in streams uncontrolled by structure, and the initial consequent streams on a coastal plain frequently diverge more widely in their courses than do the segments of northern Appalachian rivers. Obviously, considering the entire rivers including the connecting longitudinal elements of more recent origin, they are not parallel nor do they follow the shortest distances to the sea from their heads.

As a further criticism Thompson (1939, p. 1329) states:

"Although one of the main objectives of the theory of regional superposition is to explain the supposed lack of adjustment of streams to structure, Johnson notes this remarkable adjustment, which he designated modification of consequent drainage, and gives that as a reason for superposition earlier than on the Schooley peneplane."

The contradiction implied in the above sentence does not exist in Johnson's exposition of the hypothesis of regional superposition. His object was to explain the combination of two drainage elements: one transverse to structure, the other parallel to structure. In the first type adjustment is totally lacking; in the second it is almost perfectly developed. Thus both lack of adjustment and remarkable adjustment prevail simultaneously in the folded belt, but for different elements of the drainage. The hypothesis of regional superposition accounts for both elements.

Thompson (1939, p. 1330) states that alignment of water gaps and wind gaps is not common in the southern Appalachians. This is not, however, evidence against regional superposition in the folded Appalachians of Pennsylvania where alignment of gaps is conspicuous. Davis (1909) early recognized this alignment and cited it as one objection to his own interpretation of Appalachian drainage history.

The significance of steepened ridge crests adjacent to gaps has been discussed in another paper (Strahler, 1944). It was shown that the distinct sags into which water gaps have been cut would not be produced by normal noncyclic erosion of a folded belt, as Thompson (1939, p. 1330) argued. Ver Steeg (1940, p. 694) pointed out, however, that this feature of the ridge crests is not in itself evidence of stream superposition. It simply proves that the streams were in place in their water gaps at the close of the Schooley cycle, regardless of their origin. When the hypothesis of regional superposition is employed, this evidence requires that the superposition be pre-Schooley.

Von Engeln (1942, p. 359) has written that the significant difficulty in the theory of regional superposition is that, if superposition occurred on the Fall-Zone peneplane, all evidences of this transverse consequent drainage would by now have been removed. This point has been discussed in another paper (Strahler, 1944). A Fall-Zone date for superposition, far from causing difficulties, is required to explain the absence of

many wind-gap series, the form of ridge crests, and the sharp inner margin of the coastal plain.

CONCLUSION

Four hypotheses of central folded Appalachian drainage have been discussed. The hypothesis of subsequent origin of transverse Appalachian streams has been only touched upon by a number of writers. In view of the fact that growth of subsequent streams along transverse fault lines might produce the aligned groups of gaps present in the region, this explanation was studied further. Examination of water gaps of the major streams revealed no evidence that these gaps are related to transverse faults. On the other hand, in the few known instances where transverse faults break ridges, streams have not utilized the structures. Deductive analysis indicates that streams do not cut water gaps by headward growth along fault lines except under most unusual conditions.

The hypothesis of modified consequent drainage seeks to explain Appalachian drainage as having evolved continuously from a Permian consequent stream system. Modifications produced by growth of subsequent streams in weak-rock belts and in transverse fault zones, and by superposition from overturned folds and thrust sheets, have since added their characteristics to the original drainage system. Not only are the initial postulates of this hypothesis unacceptable, but the processes of superposition described are not supported by the facts and seem to oppose well-established principles of stream development. Even if the hypothesis were in all other respects valid it does not satisfactorily explain the outstanding feature of central Appalachian drainage—the presence of southeast-flowing stream segments crossing successive folds through aligned gaps cut in resistant-rock ridges. Even if one accepts the reasons which its authors give for supposing that the Permian drainage divide lay within the folded belt, a considerable drainage reversal, not discussed by them, but difficult to account for, is required.

The hypothesis of progressive piracy and local superposition involves slow inland migration of the Permian drainage divide from the crystalline rock belt across the folds to the Appalachian Plateaus. Streams that shifted the divide were postulated to have cut water gaps by headward growth through resistant rock ridges. This explanation seems inadequate because the assumed process of water-gap formation is contrary to known principles of stream development. Even if theoretically valid, the process would not reasonably account for the remarkably linear transverse elements of present Pennsylvania drainage, for the abundant development of wind gaps, for the repeated alignment of gaps in a southeast direction, or for dual water gaps maintained independently but close together in the same ridge. Field evidence indicates that drainage development in this region has been accompanied by the destruction of former water gaps, not by progressive increase in their number.

Objections to the hypothesis of regional superposition have been analyzed and are concluded to be invalid. New evidence favoring this hypothesis was found in the seeming absence of faults in the water gaps and in the fact that streams have not utilized the few transverse fault lines present in the region. Such occurrences are the expectable result of chance superposition.

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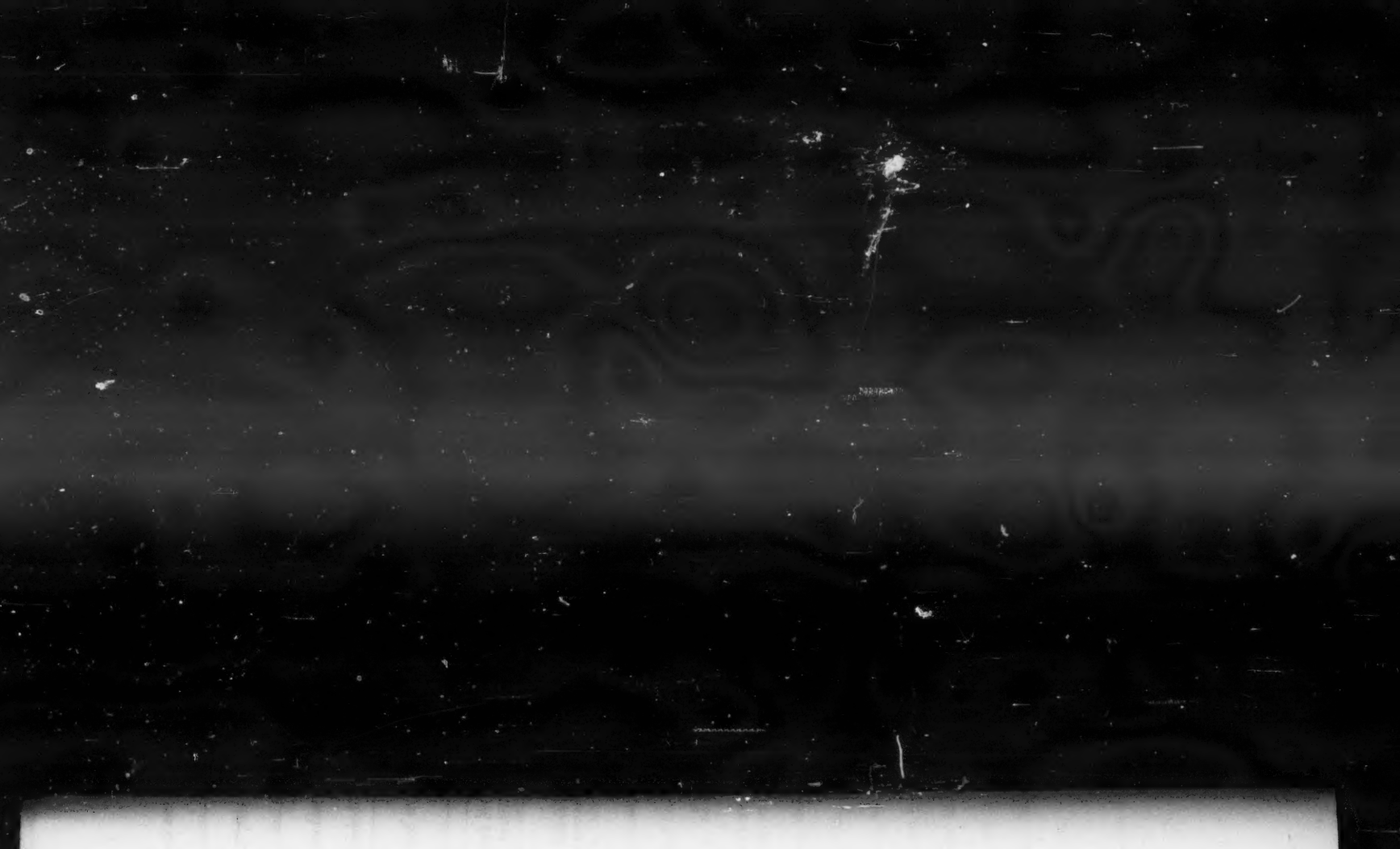

APPALAC
Ft. Mountain
Ft. Mountain
Cohutta Mtns.
Cohutta Mtns.
Rocky Face
Rocky Face
Taylor Ridge
Taylor Ridge
Lookout Mtn.
Lookout Mtn.
Sand Mtn.
Sand Mtn.
Little Mtn.
Little Mtn.
Pigeon Mtn.
Pigeon Mtn.
Johns Mtn.
Johns Mtn.
Raccoon Mtn.
Raccoon Mtn.
Walden Ridge
Walden Ridge
Cumberland Plateau
Cumberland Plateau
Cumberland Mtns.
Cumberland Mtns.
Walden Ridge
Walden Ridge
Sequatchie Valley
Sequatchie Valley
Sequatchie Valley

APPALACHIAN
PLATEAUS
GREAT
VALLEY
BALD EAGLE MOUNTAIN
NITTANY
TUSSEY MTN
BROAD MTN
KISHACOQUILLAS VALLEY
STONY MOUNTAIN
JACKS MOUNTAIN
BLUE MOUNTAIN
TERRACE MOUNTAIN
BROAD TOP MOUNTAIN
TUSCARORA MTN
KITTATINNY MTN
Juniata River
Potomac River
Raystone Branch
Eagle Creek
Lost Creek
Aughwick Creek
Tuscarora Creek
Conococheague Creek
Path Valley
Tuscarora
Jacks Mountain
Blue Mountain
Tuscarora Mountain
Kishacoquillas Valley
Nittany Valley
Broad Mtn
Tussey Mtn
Bald Eagle Mtn
Stony Mountain
Terrace Mountain
Broad Top Mountain
Kittatinny Mtn
Tuscarora Mtn
Jacks Mountain
Blue Mountain
Tuscarora Mountain
Kishacoquillas Valley
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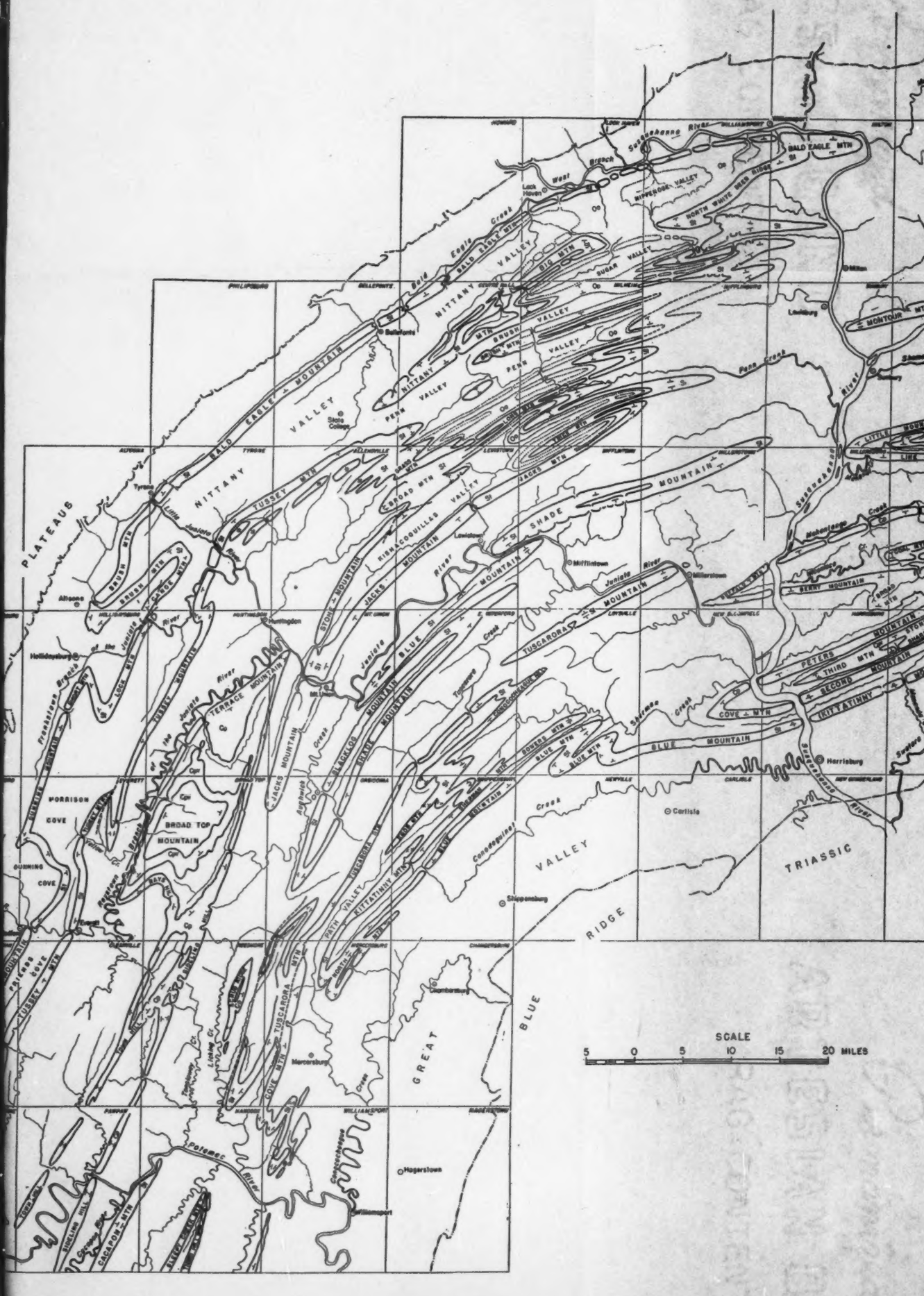
Tuscarora Mtn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GREAT VALLEY
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TRIASSIC
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JACKS MOUNTAIN
TERRACE MOUNTAIN
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SHADE MOUNTAIN
KITTATINNY MTN.
PETERS MOUNTAIN
BERRY MOUNTAIN
MONTOUR RIDGE
BIG MTN.
BROAD TOP MOUNTAIN
MORRISON COVE
COVE MTN.
Susquehanna River
Juniata River
Potomac River
Bald Eagle Creek
Penns Creek
Shermans Creek
Conodoguinet Creek
Tuscarora Creek
Lewisburg
Milton
Bellefonte
State College
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Carlisle
Harrisburg
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Chambersburg
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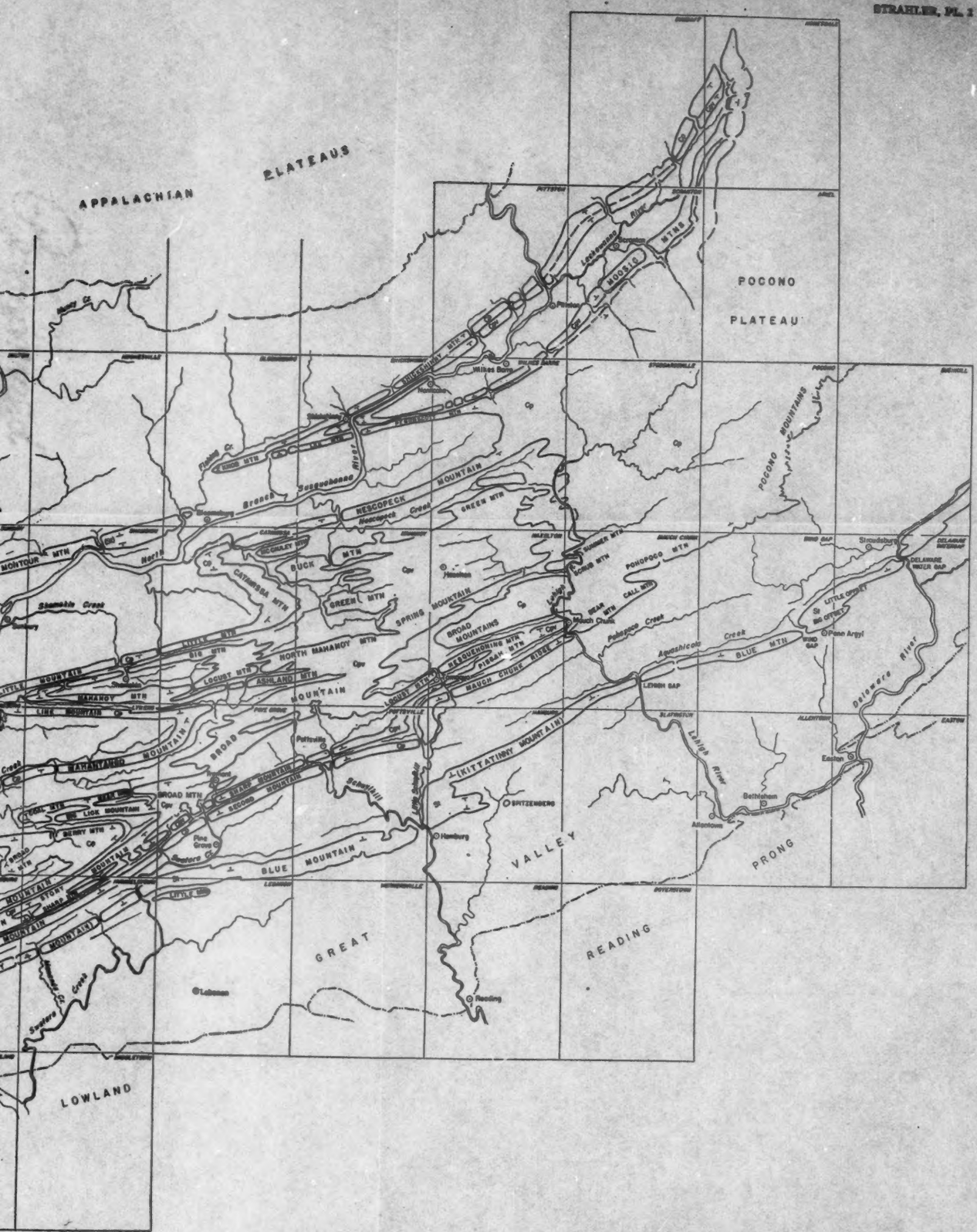
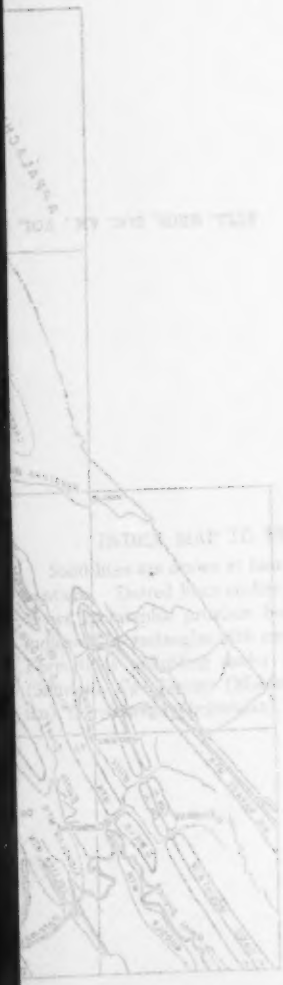


INDEX MAP TO THE FOLDED APPALACHIANS OF PENNSYLVANIA

Solid lines are drawn at base of ridges or escarpments, but do not necessarily represent geologic contacts. Dotted lines outline ridges which are only locally of major proportions. Broken line shows geomorphic province boundaries. Positions of U. S. Geological Survey quadrangles are indicated by rectangles with names in upper right corners. Structure is indicated by dip symbols. Formations upholding major ridges are designated by the following symbols: *St*—Tuscarora (Silurian), *Cp*—Pocono (Mississippian), *Cpv*—Pottsville (Pennsylvanian), *Do*—Oriskany (Devonian), *Oo*—Oswego (Ordovician).



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GLACIATION OF SIBERIA

BY RICHARD FOSTER FLINT AND HERBERT G. DORSEY, JR.

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ABSTRACT

In Siberia glaciers exist today only in arctic-maritime areas and high mountains. Former glaciers were more extensive than was generally believed prior to 1930 and had a similar distribution, closely

related to high land. Their positions were determined partly by high land and partly by the prevailing storm tracks. The moisture from which the glaciers were built came from Atlantic, Arctic, and Pacific sources and was less abundant than the moisture from which the Scandinavian ice sheet was built. The Siberian glaciers were relatively thin, but ablation was generally small.

It is believed that the present distribution of frozen ground in Siberia bears little relation to the extent of former glaciation in that region. Multiple glaciation in several parts of Siberia is established. There seems to be no good basis for the belief that the Scandinavian and Siberian ice sheets were not synchronous, although it is clear that these two glaciers reached their maxima at slightly different times.

The distribution of former glaciers is shown in two accompanying maps.

PURPOSE OF DISCUSSION

The purposes of this paper are (1) to define, for the benefit of readers outside Russia, the glaciated areas of Siberia in greater detail than was possible when Antevs (1929, p. 688) prepared his map showing the glaciation of Asia, and in particular to emphasize the relation of glaciation to topography; and (2) to discuss the manner in which the former glaciers of Siberia may have originated and to suggest a rational explanation of their growth.

ACKNOWLEDGMENT

Dr. A. L. Washburn has critically and helpfully read the manuscript.

SOURCES OF DATA

This paper is the result of a study of published texts (chiefly Russian) and maps, combined with an analysis of contemporary Siberian climates. Field study by the authors has not been possible, and, as the available Russian information on a large part of Siberia is of only reconnaissance character, the data given as to the extent of glaciation must be considered subject to revision. However, the general positions of the glaciated areas and their relations to major topographic features appear to be now well enough known so that revisions will consist chiefly of refinements in the positions of the glacial boundaries.

Until about 1930 it was the general opinion of Russian geologists that comparatively little of the vast area of Siberia had been glaciated. Thereafter the work of S. V. Obruchev and others on several expeditions to little-known parts of Arctic Siberia revealed that former glaciers did cover a large part of Siberia. At first the pendulum swung too far, and it was believed that all of Siberia north of 60° had undergone glaciation. (*See*, for example, Moltschanoff, 1926; V. A. Obruchev, 1930). Later, however, as information rapidly came in and was evaluated, the relations shown on the map (Pl. 1) became known. Figure 1 represents another opinion—that of two authors whose view is conservative in the extreme. At present it does not seem possible to combine the two concepts in a single satisfactory map.

ORIGIN OF THE FORMER GLACIERS

RELATION OF GLACIATED AREAS TO TOPOGRAPHY

A fact not adequately realized is that eastern Siberia includes large areas of mountains, some of which are very high. The steppes of western Siberia, the lowlands along the Lena[1] River, and the Arctic coastal lowlands appear to have been

[1] Transliteration of Russian place names in this paper is in accord with the usage of the Library of Congress, Washington, D. C.

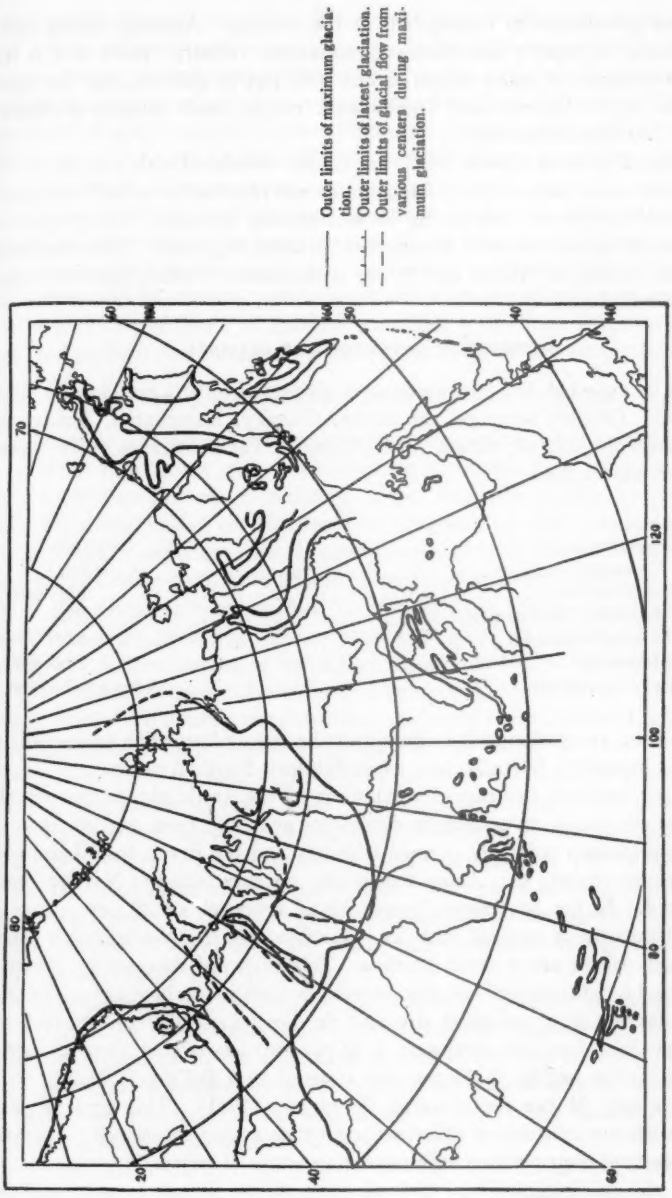
Outer limits of maximum glaciation.
Outer limits of latest glaciation.
Outer limits of glacial flow from various centers during maximum glaciation.

FIGURE 1.—*Inferred extent of former glaciers in northern Eurasia according to Gerassimov and Markov (1939)*
(*Cf.* Plate 1 for names of mountain units.)

given undue prominence in descriptions of the country. Actually almost half of eastern Siberia is properly described as mountainous country—much of it is truly rugged; the summits of many ranges exceed 5000 feet in altitude, and the tops of the massive Alpine Chersky and Verkhoyansk ranges reach extreme altitudes of 10,000 and 7000 feet, respectively.

The glaciated areas of Siberia (Pl. 1) are closely correlated with high land. The glaciated areas either are confined to mountains and plateaus or include conspicuous highlands in addition to intervening or surrounding lowlands. Every glaciated lowland area is continuous with an adjacent glaciated highland. This relationship finds a close parallel in Alaska and in the mountainous western parts of Canada and the United States.

DISTRIBUTION OF PRESENT-DAY GLACIERS

As might be expected, the few present-day glaciers of Siberia are likewise related to highlands. Glaciers occur on the Sayan, Chersky, Kamchatka, Anadyr, and Koryak mountains and on certain Arctic islands. Their altitudes (only approximately determined) are:

	Feet
Sayan Mountains	10,000
Chersky Mountains	8500
Kamchatka Mountains	5500–6000
Anadyr Mountains (south-central)	4000
Anadyr Mountains (eastern)	3000
Koryak Mountains	2700–4000
Severnaya Zemlya islands	A few hundred feet

Even from these rough figures it is clear that the regional snowline descends in the direction of increasing latitude, just as in Western North America.

Despite the relatively low rate of ablation on all the Arctic islands, the extent of present-day glaciers on these islands diminishes generally from west to east. At the west, Spitsbergen is largely covered with ice, and the Franz Josef Land archipelago is almost entirely so. About half of the northern island of Novaya Zemlya is ice-covered. In the Severnaya Zemlya island group about 40 per cent of the combined land area is covered with ice. Farther east the Novosibirskie Islands and Wrangell Island are without glaciers. This eastward decrease in glacier ice is strikingly brought out within the Severnaya Zemlya archipelago. The three principal islands are of subequal size and lie along a northwest-southeast axis. Komsomolets Island, on the northwest, is 65 per cent ice-covered, October Revolution Island, in the middle, is 45 per cent covered, and Bol'shevik Island, on the southeast, is only 21 per cent covered (Urvantsev, 1931). This decrease is not correlated with any conspicuous difference in altitude among the islands. It appears to be the result of their position with respect to source of snowfall.

Most of the precipitation observed on these Arctic islands is associated with a northeastward extension of Icelandic cyclonic activity. Dying storms, moving along the Atlantic-Arctic Front, skirt northern Scandinavia on about one fourth

of the days in the cooler months. Widespread light snowfalls normally accompany these filling cyclonic systems, which pass northern Novaya Zemlya and finally dissipate near the Novosibirskie Islands after relatively dry Arctic air has completely displaced the moisture-bearing air brought from the North Atlantic. There can be little doubt that the marked decrease from west to east in the extent of present-day glaciers on these Arctic islands is directly correlated with increasing distance from a North Atlantic moisture source.

In summary, Arctic-maritime areas appear to be the only areas capable of supporting glaciers today. The small glaciers in the continental interior are confined to the Sayan and Chersky mountains, the highest in Siberia. These exist, despite their unfavorable continental positions, probably because of relatively heavy annual snowfall and low summer temperatures prevailing at the high altitudes where they occur, and perhaps also because they occupy locally protected topographic positions.

PROBABLE ORIGIN AND GROWTH OF FORMER GLACIERS

Undoubtedly a broad regional reduction in mean annual temperature would lower the snowline in all these places, would cause the existing glaciers to expand, and would create new glaciers on mountain ranges that were uplifted at various times during the second half of the Cenozoic era, but that are slightly too low to maintain snow perennially under the climates prevailing today. If the reduction in temperature generally believed to have affected North America and Europe at the onset of each of the glacial ages be assumed to have affected Siberia as well, the general glaciation recorded by the geologic evidence apparently could have resulted. Small glaciers would have formed in the higher mountains of the interior and in much less lofty highlands along the coasts. These would have grown into piedmont glaciers which would have expanded and coalesced over intermont lowlands and minor highlands to form broadly continuous glacier covers.

The Scandinavian ice sheet, 1,000,000 to 2,000,000 square miles in area and 1300 miles in longest radius when at its maximum, has long been known to have originated on the Scandinavian Mountains as small glaciers that grew and coalesced until the mountains were nearly buried by the ice sheet thus formed. The ice sheet was fed by snowfall from moist Atlantic air masses. The great southeastward extension of the Scandinavian ice sheet was probably brought about by snowfall from cyclonic maritime air masses that skirted the polar front established by the southern border of the glacier (Flint, 1943, p. 344).

A recent discussion (Flint, 1943; *see also* Demorest, 1943) has supported the belief that the great ice sheet that covered northern North America east of the Rocky Mountains originated in the highlands of eastern Quebec, Labrador, and Baffin Island, and that, nourished by moist cyclonic air masses moving from the south and west, these glaciers expanded and thickened to form the Laurentide ice sheet, more than 3,500,000 square miles in area.

Reconstruction of the growth of an ice sheet is based on four principal points:

(1) Although precipitation today over the areas formerly covered by ice sheets is moderate to small, the air masses now traversing these regions contain abundant moisture, only part of which is precipitated. With augmented polar-front con-

ditions, such as would be brought about by growing glaciers, more moisture would be precipitated. In the preliminary period of reduced mean annual temperatures, northern areas would cool rapidly because of an increasing and persistent snow cover, while the adjacent snow-free zone to the south would experience a slow heat loss during a moist climatic regimen. The resultant quick sharpening of frontal contrasts, although providing abundant energy for vigorous cyclonic activity accompanied by heavy snowfalls in the cold sector, would not displace the storm tracks appreciably southward until the glacier-covered region had grown much larger. This increased precipitation, causing a rapid growth of glaciers, would eventually reach a maximum in the zone north of the prevailing storm tracks. Intensified anticyclonic conditions and deep cold air masses developing over the expanding ice sheet would push the polar front and storm tracks southward, temporarily establishing a new zone of maximum snowfall.

(2) Paschinger (1923) has shown that on a highland of sufficient altitude the mean annual snowfall increases upward to a maximum and then, at higher altitudes, diminishes. Under present climatic conditions the regional snowline (which is determined chiefly by summer temperatures) normally lies higher than the level of maximum snowfall. Therefore if the regional snowline descended it would successively occupy positions at which there is more snow. This would increase the persistence of fallen snow into the summer season. If the regional snowline should descend as far as the level of maximum snowfall, the development of glaciers would be favored increasingly.

(3) Brooks (1928, p. 128) has calculated that, whereas about 20 per cent of the solar radiation received per unit area by a land surface not covered by ice or snow is reflected back into space, about 80 per cent of that received by a glacier or snow-covered area is lost by reflection, and that the lost portion would be sufficient, in temperate latitudes, to melt more than 30 feet of ice over the unit area annually. Brooks inferred that this would be a large factor in the expansion of glaciers. Not only would the loss of heat have diminished melting and evaporation of ice and snow, but by decreasing the temperature of the highland it would have induced greater precipitation. When the glaciers had become extensive enough to exert a broad regional influence, their presence would have increased appreciably the thermal gradient between low latitudes and high and would thereby have increased wind velocities, atmospheric turbulence, lateral mixing of air masses, and rates of evaporation and precipitation. Once the glaciers were in existence, this could only contribute to their rapid growth.

(4) The locus of maximum snow accumulation, initially confined to the mountains on which the glaciers originated, would gradually be spread through a zone of increasing width, owing to the effect of the expansion of the glaciers into piedmont glaciers, which, being both higher and colder than the same areas under preglacial conditions, would cause increased snowfall upon them. Subsequent intensification of anticyclonic conditions over the expanding piedmonts would displace the locus of maximum snow accumulation southward or outward until the periphery of the ice sheet built from the expanded piedmonts advancing into a more temperate or drier climate, reached a position of balance between accumulation and wastage (including both ablation and discharge into deep water).

Given initially lowered temperatures and adequately moist air masses as sources of snow, these processes should have expanded the valley glaciers into piedmont glaciers and ultimately into ice sheets thick enough to bury or almost bury the mountains. Such conditions, believed to have existed in northern Europe and in northeastern North America, appear to have existed also in northwestern Siberia, where an extensive ice sheet (though probably not a very thick one) took form. In northeastern Siberia, on the other hand, the ice-sheet condition was not fully reached, owing partly to lack of sufficient atmospheric moisture and partly to high summer temperatures which caused serious losses by ablation. Finally, in southeastern Siberia dry continental conditions were so extreme that glaciers were unable to expand beyond the mountain-valley condition.

On the basis of the foregoing hypothesis of highland origin the development of the Siberian glaciers may be summarized as follows: At the beginning of each glacial age, glaciers began to form probably as a result of accumulating snowfall in mountains located near the prevailing storm tracks in Europe, Siberia, and North America. Prior to the maximum of any glacial age—that is, before glaciers began to discharge extensively into the North Atlantic Ocean and the Arctic Sea—the sources of moisture available for nourishing the growing glaciers of Siberia should have been much the same as the present sources (Pl. 1). These sources were the North Atlantic Ocean and the not-yet frozen areas of the Arctic Sea. The North Pacific Ocean supplied a limited amount of moisture to the glaciers of extreme eastern Siberia. Despite the barrier created by the Ural Mountains, Atlantic air masses penetrate today into the interior of northern Asia (although moisture coming from the south and east is baffled out by other higher mountain ranges), and such penetration should have characterized the earlier part of any glacial age. An example of the abundant precipitation at high altitudes today is Olenya Creek weather station, at 4800 feet altitude in the western Sayan Mountains. At this place the mean annual precipitation is 47 inches (Cressey, 1939, p. 114). With slightly lowered temperatures and intensified cyclonic activity highland snowfall in this region would undoubtedly have been adequate to form and maintain the glaciers recorded by the geologic evidence.

It seems probable that early in the glacial ages, as today, cyclonic storms moving eastward from the North Atlantic region frequently became regenerated in a secondary center over the Barents Sea-Novaya Zemlya-Kara Sea area. It is probable also that the Arctic front in Siberia was even more active during the summer season then, because of heightened temperature contrasts, than it is today.

The growth of the Siberian glaciers should have been rapid at first, under the combined influence of augmented polar-front conditions, a descending regional snowline, radiational heat losses, and migrating zones of maximum snow accumulation. As the Arctic Sea became choked with ice, Siberia must have been increasingly deprived of moisture from Arctic sources. With the growth of the Scandinavian ice sheet, Siberia must have undergone a reduction of nourishment from the much more important Atlantic sources. In consequence the Siberian ice sheet should have reached its maximum, and should have begun slowly to shrink, somewhat before the Scandinavian ice sheet attained its greatest extent. However, some Atlantic air undoubtedly continued to reach the Siberian glaciers and nourished

them just enough to make the shrinkage process very slow. This condition should have continued until warming of the climate in the later part of each glacial age brought about a general decline of all the glaciers of Eurasia.

Thus it seems probable that the origin and growth of the Siberian glaciers were analogous in general principle to the development of the ice sheets in Europe and North America but that the peculiarities of the Siberian glaciers, notably the relative thinness of at least some of them, were the result of a combination of climatic and topographic conditions peculiar to Siberia. Each of the facts commonly known about the former Siberian glaciers appears to be explicable on the basis of this hypothesis of highland origin. The facts as a group do not appear to be explicable on any other basis. The significant data on the distribution of former glaciers and on the relation of their distribution to climate will now be considered.

DESCRIPTION OF THE GLACIATED AREAS[2]

NORTHWESTERN SIBERIA

The boundary between European Russia and Siberia is the long north-south line of the Ural Mountains, nearly coinciding with the meridian of 60° E., and the northern continuation of this highland in the Arctic island of Novaya Zemlya. The Urals reach altitudes of about 5000 feet, although throughout most of their extent their crest is somewhat lower. The high northern part of Novaya Zemlya, at present covered by an ice sheet, reaches an extreme altitude of nearly 3500 feet.

East of the 90th meridian, and separated from the Urals by the broad Ob River lowland, is a series of three more highlands: on the south the Putorana Mountains (5000 feet), in the middle the Byrranga Ridge (3500 feet) forming the Taimyr Peninsula, and on the north the Severnaya Zemlya archipelago (1500 feet) in the Arctic Sea.

When glaciation was most extensive (the Third Glacial Age according to recent Russian authorities) this entire group of highlands—Urals, Novaya Zemlya, Putorana, Byrranga, and Severnaya Zemlya—, together with the lowlands between them, constituted a single glaciated area of more than 1,300,000 square miles. Gerasimov and Markov (1939, p. 445) refer to the combined glacier as the Siberian ice sheet. Measured on the flank of the Urals this ice mass was about 2300 feet thick, compared with estimates of 6600 to 10,000 feet for the thickness of the Scandinavian ice sheet over central Sweden. Nowhere was it quite thick enough to bury the highlands.[3]

Geologic evidence indicates that glaciers formed independently in each highland, spread outward, and coalesced to form the Siberian ice sheet. West of the Urals this ice merged with the northeastern edge of the Scandinavian ice sheet. Even the Scandinavian ice sheet was thin in this vicinity because snowfall was much less here than on the Scandinavian Peninsula. The Siberian ice sheet was fed by

[2] This description is drawn largely from two sources (Gorkin, 1937; Gerasimov and Markov, 1939). Where specific references are not given, the authority of one or both of these two sources is implied.

[3] Gerasimov and Markov (1939) believe that the glacier had no convex-up surface independent of these highlands. Whether or not this is true (there appear to be mechanical difficulties in the way of accepting it fully) the genetic relation of this vast ice mass to the highlands is clear.

snowfall from moist Atlantic air masses. Initially this maritime air reached the highland group by passing to the north and south of Scandinavia and brought with it adequate moisture. However, with the growth of the Scandinavian ice sheet, choking of the Arctic Sea with floating ice and southward displacement of storm tracks must have reduced greatly the nourishment of the Siberian ice sheet. Yet some Atlantic air undoubtedly continued to reach western Siberia via the routes both north and south of Scandinavia without prior forced ascent to heights much above sea level. The Siberian highland group now receives a greater annual snowfall than any other comparable area in the U.S.S.R.

The relatively favorable position of the Scandinavian ice sheet with respect to snowfall enabled it to expand somewhat into the area west of the Urals vacated by the Siberian ice sheet when the latter began to wane during the last glaciation (Ramsay, 1912). This event was similar in some respects to the sequence of movements from at least two different centers of glacial outflow that affected northern Minnesota at about the same time.

During the latest glacial age the Siberian ice sheet, like its neighbor to the west, was much less extensive than during the Third Glacial Age. Although ice formed and flowed radially outward from each highland area, the ice from the Urals and Novaya Zemlya reached southward only to Latitude 64. On the east it reached the Enesei River only near the Arctic coast, where it appears to have coalesced with the Severnaya Zemlya-Byrranga ice mass, then flowing westward (Urvantsev, 1931). At this time the Putorana Mountains constituted a separate isolated center of glaciation.

CENTRAL SIBERIAN PLATEAU

The vast rolling region between the Enesei River and the Lena River is the Central Siberian Plateau. The summits of its hills reach altitudes of 2000 to 2500 feet throughout most of its extent. However, a broad east-west swell at the latitude of the Arctic Circle rises to more than 3500 feet. This highland constituted a separate center of glaciation that expanded to more than 200 miles in greatest length. The date of the glaciation is uncertain. Presumably the ice reached its greatest extent during the Third Glacial Age, but whether there was any ice here during the latest glacial age is not known. The ice probably formed a thin ice cap like that which covers northern Novaya Zemlya today. Cyclonic storms providing nourishment for this center of glaciation must have come principally from the North Atlantic by the same routes as those that reached the Siberian ice sheet. The small quantity of ice here was undoubtedly the result of the limited amount of moisture available so far from its source.

In the southwestern part of the Central Siberian Plateau, along the eastern side of the Enesei River valley at about Latitude 60 another swell, trending north-south, is known as the Enesei Hills. Small summit areas, 3500 feet in altitude, may have been the sites of local glaciers, according to Russian data. If they were, the glaciers were probably contemporaneous with the greatest southeastward extent of the Siberian ice sheet, which at one time reached to within 70 miles of the Enesei Hills and induced polar-front climatic conditions in the surrounding country. It seems

unlikely that this area supported glaciers during the latest glacial age when the nearest margin of the Siberian ice sheet was 450 miles to the northwest.

MOUNTAINS OF NORTHEASTERN SIBERIA

General glacial relations.—The northeastern part of Siberia east of the 125th meridian is marked by a vast system of arcuate mountains trending in various directions. Some of the ranges are high, with Alpine summits. From west to east the principal units and their extreme altitudes are (Pl. 1):

Verkhoyansk Mountains (7000 feet)
Chersky Mountains (several ranges) (10,200 feet)
Kongin Mountains (7000 feet)
Kolyma (Gydan) Mountains (7000 feet)
Aniui Mountains (6000 feet)
Anadyr Mountains (7000 feet)
Koryak Mountains (6000 feet)
Kamchatka Mountains (8000 feet, with single peaks up to more than 15,000 feet)

These mountains supported a complex system of valley, piedmont, and ice-cap glaciers which as a whole was much like the former Cordilleran glacier complex of western North America and was similar (on a much larger scale) to the system of glaciers that exists today between Mt. St. Elias and Mt. Logan in coastal Alaska.

The glaciers flowed radially outward from each mountain range named above and from subsidiary ranges, and at their maximum extent coalesced into an almost continuous system that extended from the Lena River on the west to Bering Strait on the east, and from the Arctic Sea southward to the Sea of Okhotsk. This nearly coalescent system was 1800 miles long—almost exactly the length of the continuous part of the North American Cordilleran glacier complex measured from the State of Washington to western Alaska. The area involved was close to 1,000,000 square miles.

In addition to the main coalescent mass there were two independent areas of glaciation—the Koryak Mountains on the Bering Sea coast and the high volcanic chain on the Kamchatka Peninsula.

Throughout this entire complex little evidence of the dates of glaciation has been gathered. The glaciers had their greatest extent during a glacial age (presumably the Third) that antedated the latest age, but differentiation of successive glaciations has been made in only a few localities. According to Obruchev and Salischev (1935, p. 640) there is evidence in the Verkhoyansk Mountains of three glacial stages and in the region of the Anadyr Mountains of two glacial stages. In general the latest glaciation, like that in northwestern Siberia, was far less extensive than its predecessor and was confined to the higher parts of the higher mountain ranges. Except in the Koryak and Kamchatka areas, the moisture available for nourishment was less than that supplied to the Siberian ice sheet.

Maritime region.—In the maritime region that constitutes the southeastern coastal part of the mountain system of northeastern Siberia glaciation reached the sea extensively. At the head of the Sea of Okhotsk, along the Pacific coast of Kamchatka Peninsula, along the coast of Bering Sea and Bering Strait, and along the Arctic

coast north of the Anadyr Mountains, the former glaciers reached what is now sea level and extended out to unknown distances over what is now the continental shelf. Both the Kamchatka and Koryak glaciated areas extend to low altitudes. This fact implies that the maritime positions of these two highland masses resulted in their receiving relatively heavy snowfall.

In southern Kamchatka the glaciated area reaches the sea on the east side of the peninsula, but on the west side it fails by almost 1000 feet to reach sea level. This discrepancy suggests that during the glaciation, as today, a strong difference existed between the moist eastern slopes and the less moist western slopes of the Kamchatka peninsula.

The Arctic coast is a cold-maritime region. The glaciation of this region, centering in the Anadyr and Aniui mountains, apparently reached its rather large extent as a result less of heavy snowfall than of low summer temperatures. The present-day precipitation of this region is only one-third to half that on Kamchatka. The former glacier cover was probably thin as compared with the glacier ice farther south and west.

Verkhoyansk and Chersky mountains.—The massive Verkhoyansk and Chersky mountain arcs are the highest and most continuous mountain chains of northeastern Siberia. The two masses are separated by a great basin 200 miles wide, drained by the Yana River. On the outer slopes of the inclosing mountains—that is, the west slope of the Verkhoyansk Mountains and the northeast slope of the Chersky Mountains—the former glaciers were valley glaciers that were confined to the mountains. On the other hand the glaciers of the inner slopes flowed into the Yana River basin, coalesced, and eventually filled the southern part of the basin nearly to the brim. This great accumulation of ice, 2300 feet thick, is attributable in part to the facts that glaciers were fed into the basin continuously from around three sides and that the basin narrows toward its outlet at the north. Probably far more significant, however, was the fact that the basin was frigid, being shielded on the west, south, and east by mountains at least 7000 feet high, so that ablation was held to a minimum. The geologic evidence indicates that at the glacial maximum virtual ice-sheet conditions were established over the entire basin, while valley glaciers continued to drain the outer slopes of the inclosing ranges. The relations thus pictured are closely similar to those which characterized southern British Columbia. There the "Interior Plateau" between the Coast Ranges on the west and the high Rocky Mountains on the east was filled with ice originally derived from the high mountains but for a time approaching or reaching the condition of an ice sheet. Yet the glaciers that drained the outer flanks of both mountain chains were of the valley-glacier type (Daly, 1912, p. 577).

Kongin and Kolyma (Gydan) mountains.—The intermont ice-cap condition did not exist to the same extent elsewhere in northeastern Siberia because the Chersky and Verkhoyansk mountains lie much farther apart than do any of the other ranges. During the glacial maximum the intermont lowlands between the various ranges of the Chersky Mountains and between the Chersky, Kongin, and Kolyma (Gydan) mountains were filled with ice which was continuous across at least the lower parts of the intervening ranges.

On their southern slopes, however, all these ranges were drained by valley glaciers. These reached down to progressively lower altitudes toward the sea, in response to increased snowfall and decreased summer temperatures in the vicinity of the coast. At the south end of the Kolyma (Gydan) Mountains the glaciers are said to have reached what is now sea level. Likewise the eastern flank of this same mountain range was drained by valley glaciers which under the prevailing maritime climate descended to an altitude of less than 700 feet.

Chukotsk Peninsula and Arctic coast.—It is probable that the glaciation of the large peninsula occupied by the Anadyr Mountains spread outward not only from centers in these mountains but also in part from centers in other, lower mountain units, the ice from the several sources coalescing to cover virtually the entire peninsula.

The wide area along the Arctic coast in the region of the lower Indigirka and Kolyma rivers was not glaciated because precipitation is small and the region is too low to retain winter snowfall throughout the summer. Glacier ice flowing from the high Chersky Mountains to the south did not reach this lowland because the bulk of the snowfall from moist air masses moving from the west and south was precipitated mainly on the western and southern sides of these mountains. The starved condition of this part of the Siberian Arctic coast is roughly analogous to that of the Arctic coast of Alaska, which was little glaciated because the high Brooks Range hedged off the Pacific air masses—its best potential source of moisture.

It is important to repeat that the conditions described for northwestern Siberia are conditions that existed during the maximum glaciation, probably at the time of the Third Glacial Age. During the latest glacial age the glaciers were much less extensive, though their limits have not yet been mapped.

ISLANDS IN THE ARCTIC SEA

All the major island groups in the Arctic Sea—Spitsbergen, Franz Josef Land, Novaya Zemlya, Severnaya Zemlya, the Novosibirskie Islands, and Wrangell Island—have been completely glaciated at one or more times during the Ice Age. The first two are outside the scope of the present paper and will not be discussed here; we have no evidence as to the outer limits of their glaciers beyond the island shores so that we do not know whether or not they were ever coalescent either with each other or with the main Scandinavian ice sheet.

During the maximum glaciation, glaciers that radiated from Novaya Zemlya and Severnaya Zemlya coalesced with glaciers originating in centers on the mainland to form the Siberian ice sheet. Coalescence was possible because the intervening water areas are shallow. The sea floor, whatever its degree of emergence may have been at that time, differed little in general character from the low plains along the Arctic coast today.

Although the Novosibirskie Islands have a maximum altitude of barely 700 feet, they were the center of a small ice sheet that attained a radius of more than 150 miles and expanded across the shallow sea floor to the low mainland coast between the mouths of the Yana and Indigirka rivers. Apparently in this region of comparatively little snowfall nourishment was inadequate, even at the glacial maximum, to permit coalescence of this ice sheet with the glaciers that flowed north out of the

Chersky Mountains. Ermolaev (1932) states that the ice was always thin. Elsewhere than in this mainland sector the outer limits of the Novosibirskie Islands ice sheet are conjectural.

Wrangell Island, with a maximum altitude of 2000 feet, was the center of an ice cap, similar to the Novosibirskie Islands ice cap but smaller. The Wrangell Island ice failed to reach the mainland, though the intervening distance, across shallow water, is little more than 100 miles. At its maximum this ice sheet completely blanketed Wrangell Island; nothing further is known regarding its limits. Because the position of the island is unfavorable for the receipt of snowfall, it is not likely that the ice sheet was extensive.

HIGHLANDS OF THE AMUR-LENA REGION

Dzhugdzhur Mountains.—The country between the Amur and Lena rivers, east of the Lake Baikal region, includes a number of mountain masses several of which (the highest ones in the region) formerly generated glaciers. One of these was the Dzhugdzhur Mountains, which form the northwest coast of the Sea of Okhotsk. Although this highland reaches little more than 4000 feet altitude, it supported glaciers throughout a distance of 250 miles measured along the crest of the range. Here much orographic snow must have been precipitated out of moist easterly winds blowing from the Sea of Okhotsk.

Stanovoi Mountains and adjacent highlands.—The long Stanovoi Range extends westward from the Sea of Okhotsk along the 56th parallel. Well above 4000 feet throughout most of its length, the crest in two places reaches altitudes of 7500 feet. The range is said to be glaciated throughout a linear distance of about 450 miles, the glaciation reaching lower altitudes with increasing proximity to the sea. Although moisture from the Sea of Okhotsk supplied the principal nourishment, some cyclonic precipitation probably reached the Stanovoi Range from the southwest.

The northern base of the Stanovoi Mountains merges into the Aldan Plateau, which, near Latitude 58, rises to form two high areas respectively about 6500 and 8500 feet above sea level. Both areas were the sites of local glaciers. These were fed principally by the snowfall, at high altitudes, associated with storms moving from the west during the warmer months along an intensified polar front that does not exist here today.

South of the eastern end of the Stanovoi Mountains a similar highland (the Yamalin Range), reaching an extreme altitude of 7300 feet, is believed to have carried small local glaciers. These would have received nourishment from disturbances moving eastward along the polar front, in addition to moisture from cyclonic northeasterlies blowing from the Sea of Okhotsk.

HIGHLANDS OF THE BAIKAL REGION

Lake Baikal lies in the midst of a complex of high mountains and plateaus between the headwaters of the Amur and Lena river systems. The principal elements in this complex are the Baikal Mountains, bordering the lake on the west, the Baikal-Olekma highland (comprising several individual ranges) northeast of the lake, and the Vitim Plateau east of the lake. All these highlands are rugged, with barren

and in places serrate crests. Throughout much of their extent they stand well above 5000 feet and in places reach extreme altitudes of 10,000 feet. They are the sites of a former complex of glaciers which when at their maximum extent are believed to have been coalescent. According to Eskola (1929) glacier ice near the head of Lake Baikal was 1700 feet thick. The topography of the highlands and the general character of the glaciation are like those of the Yellowstone Park district in western United States.

South of the Vitim Plateau the Yablonovi Mountains, with a maximum altitude of 5500 feet, trend northeast. They are reported to be glaciated throughout a distance of 500 miles. A minor highland south of the western end of the Yablonovi Mountains is believed also to have been glaciated.

Probably all these highland glaciers were fed by snowfall from storms moving eastward along a well-defined polar front that existed during the summer at the times of maximum glaciation.

ALTAI HIGHLANDS

The great Altai mountain system in the border region of Russia and Mongolia includes a number of ranges, of which the Sayan Range, whose rugged crest stands in most places above 7500 feet and at one point reaches nearly 11,500 feet, lies within Siberia. According to Granö (1910, p. 219) the Sayan was the site of a glacier complex 200 miles or more in length and in some places more than 60 miles wide, with glaciers reaching down to altitudes of less than 3000 feet. This complex must have resembled the former glacier systems of the Sierra Nevada and Cascade highlands in western United States. The Sayan glaciers probably derived their nourishment from cyclonic activity along a pronounced polar front. Such a front would have resulted from a southward shift of the present Siberian Arctic Front that develops today during the summer season, hundreds of miles to the north.

SUPPOSED RELATION OF FROZEN GROUND TO FORMER GLACIATION

No discussion of the former glaciation of Siberia would be complete without mention of the old but still-persistent hypothesis that areas of perennially frozen ground in Siberia are complementary to the areas that were formerly glaciated. (*See* Nikiforoff, 1928; Cressey, 1939.) This view originated in the nineteenth century but is still quoted. (*Cf.* Gerasimov and Markov, 1939, p. 446.) Apparently the argument has never been stated fully and clearly, but in general it is as follows:

(1) There is much perennially frozen ground in Siberia.

(2) Frozen into it have been found carcasses of the woolly mammoth (*Mammonteus primigenius*) and the woolly rhinoceros (*Coelodonta*), which are believed to have become extinct either in late-preglacial time or at any rate no later than the last glacial age.

(3) Therefore the freezing is believed to date from the last glacial age or earlier.

(4) The ground could not have become frozen it it had been covered by a thick blanket of glacier ice. Hence the glaciation of Siberia must have been confined to those areas which are not now frozen.

It is quite true that freezing of the ground has resulted from the abstraction of heat from the ground during the long and extremely cold Siberian winters. It is probably true also that the ground would not freeze deeply, if at all, if it were covered by an ice sheet. However, though the hypothesis just outlined may have sufficed when knowledge of the extent of both glaciation and frozen ground was very scanty, it does not meet the facts known today.

The areal limits of frozen ground and of former glaciation are now fairly well known. They are not complementary. In western Siberia the glaciated area extends far south of the southern limit of frozen ground, whereas southeastern Siberia, much of which was never glaciated, is widely underlain by frozen ground.

Furthermore since no evidence has been adduced to prove that mammoths and rhinoceroses did not persist into post-ice-sheet time (as mammoths clearly did in North America) the remains of these animals in frozen ground proves nothing regarding the date of freezing of the ground.

The growing opinion among Russian geologists is that the ground now frozen has reached that condition since the last glacial age, in direct response to the existing climate. Whether or not this opinion is confirmed in future, the hypothesis that from the areal distribution of frozen ground much can be inferred as to the extent of former glaciation in Siberia is untenable.

MULTIPLE GLACIATION

Evidence of multiple glaciation in Siberia is clear and has been reported by many geologists in localities from the Ural Mountains to the Bering Sea. In the region of the southern margin of the former Siberian ice sheet two glacial stages (correlated with the Riss and Würm) are reported to be separated from each other by sediments with interglacial flora and fauna (Vvedensky, 1933). In Severnaya Zemlya two stages are recorded (Kalessnik and Bykova, 1937). In the Novosibirskie Islands there are three stages separated by marine deposits (Ermolaev, 1932). In the Verkhoyansk-Chersky mountains region S. V. Obruchev (1933) has observed at least three stages, and, in the Chukotsk Peninsula, two stages, of which the later is identified by him (1939) with the Wisconsin stage in Alaska. Two stages have been recognized on the Kamchatka peninsula (Gerasimov and Markov, 1939, p. 449). Three stages are recorded in the Kolyma Mountains, two in the highlands around Lake Baikal (Denguin, 1930; Bobin, 1933; Pavlovski, 1933; Pilipenko, 1934; Dumitrachko, 1939), and at least two in the Sayan Mountains (Granö, 1910). In the mountain regions of eastern Siberia evidence of multiple glaciation is said to consist of end moraines and terraces rather than of fossil-bearing interglacial deposits between till sheets; hence its reliability must be considered as uncertain.

Although correlations are still tentative, the multiple glaciation of Siberia is firmly established. Probably the several glacial ages were synchronous, but this is not universally admitted. Gerasimov and Markov (1939, p. 447, 449) believe that the glaciers of dry continental Siberia could not have existed at the same time as the Scandinavian ice sheet in maritime Europe because the development of an ice sheet in Europe would shut Siberia off from what little precipitation it now receives.

However, this view does not appear to be warranted. Early in any glacial age the preglacial storm tracks would not have been displaced southward to a significant extent, and in consequence abundant nourishment should have been available for Siberian glaciers. Furthermore the secondary center of cyclonic activity located in the Barents Sea-Novaya Zemlya-Kara Sea area should have been intensified, at least temporarily, by increased frontal contrasts, especially in the warmer months. Furthermore, although probably the Siberian anticyclone would have inhibited winter precipitation to a greater extent than at present, it should have given way to continental heating in summer with accompanying strong cyclonic developments along the Siberian Arctic front. These conditions undoubtedly combined to provide a sharp increase in total precipitation amounts over Siberia, thus causing a relatively rapid growth of glaciers in the highland areas at the same time as in the Scandinavian mountains.

Subsequently, as the glaciers coalesced and the Siberian and Scandinavian ice sheets took form, anticyclonic circulations were locally induced over them. There is little doubt that then (but not until then) Siberia was increasingly shut off from Atlantic moisture. This deduction is consistent with the relative thinness and restricted extent of the Siberian glaciation as compared with that of the Scandinavian center.

Thus, there seems to be no sound basis for the belief that the Scandinavian ice sheet and the Siberian glaciers were not synchronous. Furthermore it is very improbable that any climatic change causing renewed glaciation of Siberia would not reconstitute the Scandinavian ice sheet. Conversely, when (as at present) the temperature is too high to permit the maintenance of an ice sheet in Scandinavia, there is no chance of large glaciers forming in western Siberia with its high summer temperatures.

EXTENSIVE GLACIAL LAKES AND CHANGES OF LEVEL

Matters such as changes of level and glacial drainage in Siberia are not yet sufficiently well known to justify extensive discussion. In general, however, emerged marine deposits identified with both glacial and interglacial ages exist along both Arctic and Pacific coasts, in places extending more than 200 miles inland. (*See* especially V. A. Obruchev, 1935–1938, p. 1211). Undoubtedly the emergence of these deposits is the combined result of crustal warping and fluctuation of sea level. Further, the region fringing the Siberian ice sheet on the south, between the Urals and the Enesei, was the site of an extraglacial lake or lakes (Vvedensky, 1933; Moltschanoff, 1926). The impounded water, held by a broad glacier dam in the drainage basins of the Ob and the Irtysh, seems to have been controlled by a spillway at the head of the Tobol River (Lat. 65°E.; Long. 52°N.) that discharged into a route now dry leading to the Aral "Sea" east of the Caspian. Probably the development of this lake or lakes paralleled the sequences of lakes formed in Fennoscandia and in central North America during the shrinking of the last ice sheets in those regions.

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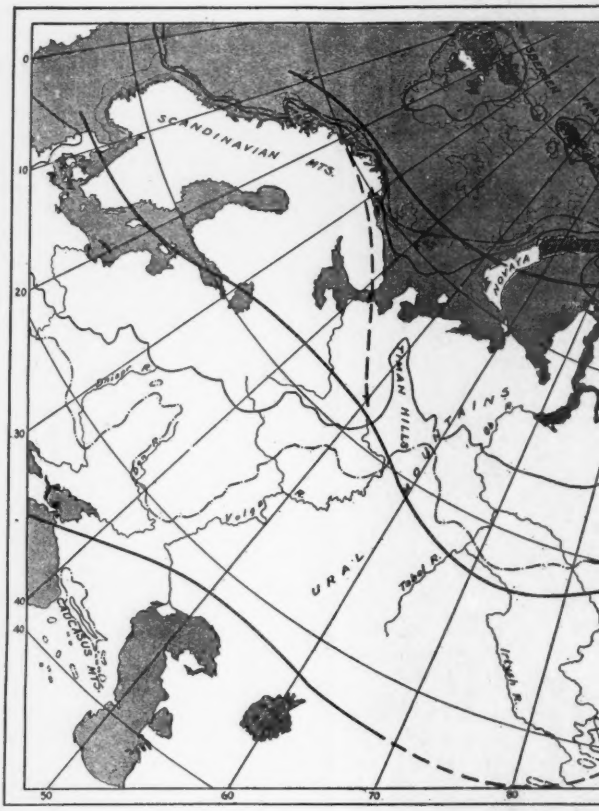


SKETCH MA

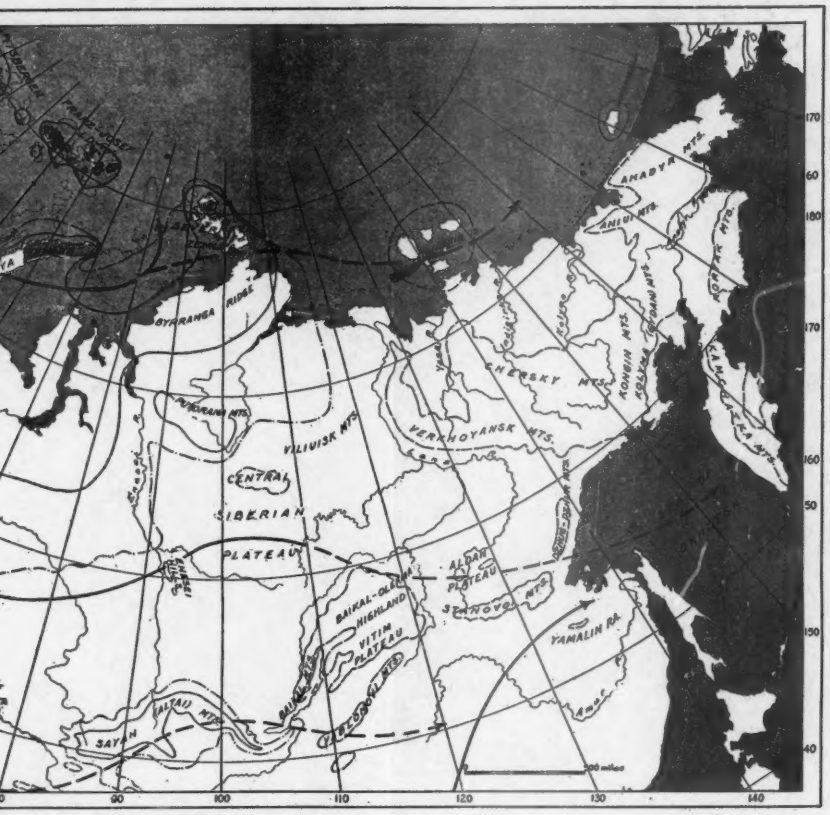
Showing existing glaciers, inferred extent of former glaciers (mainly acco

Coarse stippling: Existing glaciers large enough to show to scale.

Major present-day storm tracks (annual mean).

Occasional and seasonally important present-day storm tracks (annual mean).

Outer limits o
tion (latest gla
in Europe and
Siberia.



ETCH MAP OF NORTHERN EURASIA

ainly according to the Soviet Atlas—Gorkin, 1937), principal mountain units, and storm tracks.

ter limits of glacia- n (latest glacial age) Europe and western eria.

Outer limits of glaciation in Europe and western Siberia when glaciers were most extensive (regardless of age).

Outer limits of glaciation in eastern Siberia (age undetermined; include some areas of doubtful glaciation).

200-meter depth line offshore.